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A Guidebook to

THE GEOLOGY OF NORTHEASTERN MAINE AND NEIGHBORING NEW BRUNSWICK

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The 72nd Annual Meeting of the
New England Intercollegiate Geological Conference

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New Hampshire
Library*

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NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

Informally organized in 1901 by William M. Davis, the NEIGC provides an opportunity for students, faculty, and industrial geologists to share the results of recent geologic studies in New England. With the exception of eight years during the two World Wars, the conference has met every fall since the original trip to the Connecticut Valley led by Prof. Davis. The 1980 meeting in Presque Isle is the 72nd one held by the conference and the tenth meeting in Maine. It was not until 1925 (in Waterville) that the conference held its first meeting in Maine and the state was revisited about every ten years until 1960 (1934, Lewiston; 1950, Bangor; 1960, Rumford). Since 1960 the conference has been returning to Maine more frequently which attests to the increasing amount of new work being done in New England's largest state and the greater ease with which people can travel to and within the state (1965, Brunswick; 1966, Katahdin; 1970, Rangeley; 1974, Orono; 1978, Calais). This year's meeting is the first in Aroostook County, Maine's largest, and it is fitting that trips in neighboring New Brunswick are included since the northwestern part of the province has many geological and cultural ties to "The County."

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The editors wish to thank Marjorie H. Roy for her assistance in the final assembly of this Guidebook.

IN DEDICATION



Ely Mencher
1913 - 1978

This guidebook is dedicated to Ely Mencher who was a friend, teacher, and colleague to over a generation of students at MIT and later at City College. His work from 1962 to 1976 and those of his undergraduate and graduate students in northern Maine established the foundation upon which much of this conference is based. His quiet wit, generosity, and sense of responsibility endeared him to those of us who had the privilege to meet and work with him.

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TECTONICS AND SEDIMENTATION IN NORTHEASTERN
MAINE AND ADJACENT NEW BRUNSWICK

by

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The field trips of the 1980 NEIGC will examine the bedrock and surficial geology of a large region in northeastern Maine and neighboring New Brunswick as shown in Figure 1. The history of intensive geologic studies in this region is more recent than for more southerly parts of New England but the results of the studies are important to the "big picture" because the major rock belts within the central part of the northern Appalachian range pass through it and are seen at lower metamorphic grade. In fact, the best fossil-dated stratigraphic cross-section of the range from the Connecticut Valley-Gaspé Synclinorium through the Bronson Hill-Boundary Mountain Anticlinorium and across the western flank of the Merrimack Synclinorium is present in northern Maine and western New Brunswick.

The bedrock trips during the conference will not show the complete section but will give participants a clear view of the complex stratigraphy from the Bronson Hill-Boundary Mountain terrain across the Aroostook-Matapedia Belt, to the Miramichi Belt of northern New Brunswick. The essential features of the stratigraphy of the region are summarized in Figure 2. As will be illustrated below, the early Paleozoic stratigraphy of the region of this year's conference is similar to that seen at higher metamorphic grade during the 1970 Rangeley conference and more recently during the 1974 conference in Orono. The purpose of this article is to provide an overview of the regional stratigraphy in the region covered by the conference. In addition an attempt is also made to show both the interplay of tectonics and sedimentation suggested by the stratigraphic variations and to illustrate the broad regional continuity of the resulting tectono-stratigraphic picture. It is hoped that the treatment will help participants to keep track of the formational "players" in the game and to begin to see the positions they may have played during the evolution of this part of the northern Appalachians.

Pre-Middle (pre-Caradocian) Ordovician Stratigraphy

In northern Maine our view of the pre-Middle Ordovician is very restricted. A sparsely fossiliferous but relatively thick pre-Caradocian section has been described by Neuman (1967; Neuman and Rankin, this volume) in the core of the Weeksboro-Lunksoos Lake Anticline. These rocks include the Grand Pitch Formation (variegated slate and quartzite) and the overlying Shin Brook Formation (tuffaceous volcanic rocks and minor sedimentary rocks). Rocks similar to the Late Precambrian/Early Cambrian Grand Pitch have been mapped to the northwest of the Weeksboro-Lunksoos Lake anticline by Hall (1970) as the Chase Brook Formation. The lower quartzite-rich unit of the Tetagouche Group of the Miramichi Anticlinorium in New Brunswick is commonly considered to also be equivalent to the Grand Pitch (Rast, St. Peter, and Lutes, this volume).

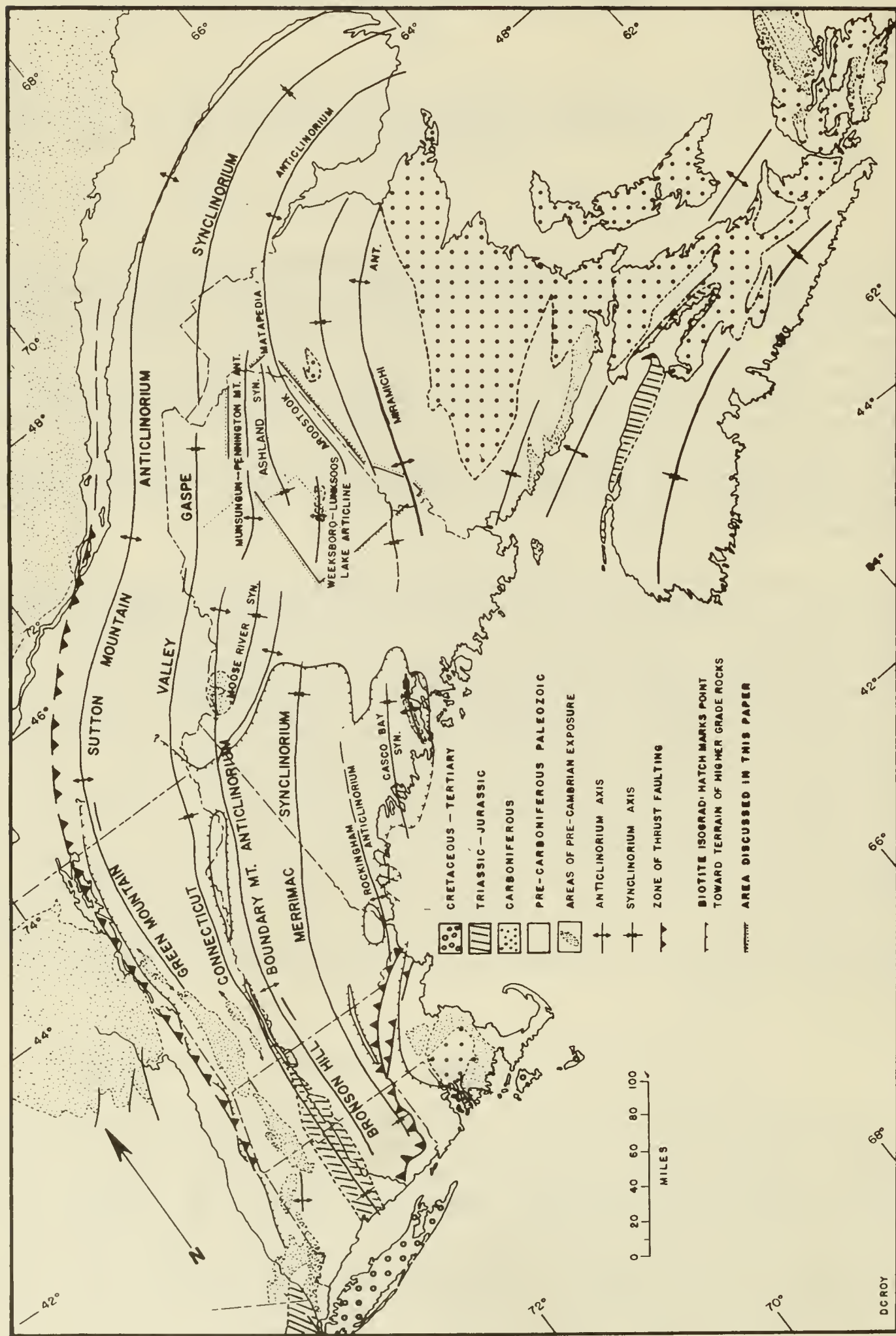


Figure 1: Principal tectonic features of the Northern Appalachians in New England, Eastern Quebec, and New Brunswick.

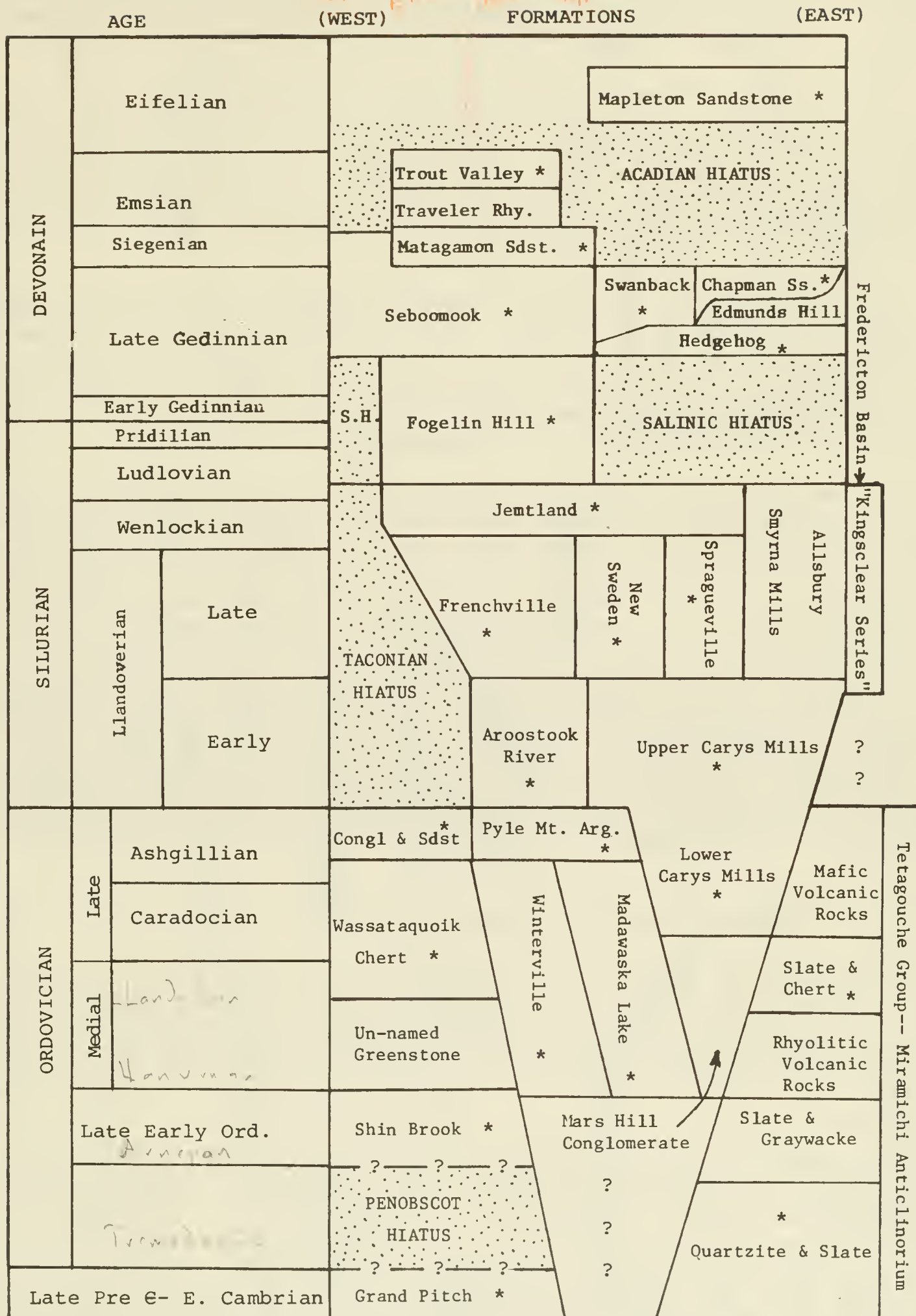


Figure 2: Stratigraphy of northeastern Maine and adjacent parts of New Brunswick. Star indicates that the unit is dated by fossils.

The Miramachi belt is, however, separated from the Weeksboro-Lunksoos Lake Anticline by younger rocks of the Aroostook-Matapedia Belt and it is therefore difficult to establish lateral continuity of the quartzite-rich sequences.

The Shin Brook Formation overlies the Grand Pitch unconformably and contains Late Early or Early Middle brachiopods and trilobites (Neuman, 1967). No clear lithologic correlatives of the Shin Brook have been described in Maine or New Brunswick although largely slate-graywacke sections elsewhere are probable temporal equivalents (Neuman, 1967; Hall, 1969). The slate-siltstone-graywacke unit of the Tetagouche Group (Rast, St. Peter, and Lutes, this volume) may be in part correlative with the Shin Pond.

The unconformity at the base of the Shin Pond Formation may represent a major hiatus and has been attributed to the Penobscott Disturbance by Neuman and Rankin (1966). An unconformity in apparently similar stratigraphic position may be widespread in the northern Appalachians (Hall, 1969), but good dating of the unconformity is not everywhere possible and the deformation and erosion suggested by it may not represent a widespread synchronous event.

The paleogeography of the earliest Paleozoic is not well understood as might be expected from the limited distribution of these poorly dated rocks of pre-Middle Ordovician age. Since these rocks are found in anticlinorial tracts that are separated by synclinorial belts of younger rocks it is difficult to access regional patterns. Reliance on lithologic correlations, especially across the regional structure, is dangerous because rapid cross-strike facies changes are common in younger rocks (Roy and Mencher, 1976) and are likely in the pre-Middle Ordovician as well. It seems clear that the Grand Pitch-type lithofacies in which abundant quartzose sandstone is interlayered with slate (in variable proportions) is widespread in the northern Appalachians and probably underlies much of the region of this conference. The abundance of quartzite in the lithofacies suggests sialic (cratonal?) sources along the margins of and possibly within the orogen. The variations in abundance of quartzite may reflect changes in proximity to the sialic sources.

Late Ordovician (Caradocian/Ashgillian) Lithofacies

Beginning with rocks of the later part of the Ordovician it becomes possible to work out fairly clear lithofacies patterns in northeastern Maine and parts of New Brunswick. Such a lithofacies pattern involving three major lithofacies and representing the Caradocian and Ashgillian is shown in Figure 3.¹

¹In figures 3,4,5, and 6 the following information applies. Towns and cities are abbreviated as follows: A, Ashland; Aug, Augusta; B, Bathurst; Ban, Bangor; C, Caribou; Cal, Calais; D, Dalhousie; F, Fredericton; FK, Fort Kent; G, Greenville; GF, Grand Falls; H, Houlton; MT. K., Mount Katahdin; M, Matapedia; P, Portage; PI, Presque Isle; R, Rockland; VB, Van Buren. Primary sources of information are: Roy, 1970; Roy and Mencher, 1976; Pavlides, 1965, 1968, 1971, 1972, 1973; and Hall, 1970; Neuman, 1967; Boudette and others, 1976; St. Julien and Hubert, 1975; Boudette and Boone, 1976; Boone, 1973; Anderson, 1968; Hamilton-Smith, 1969, 1970; Boucot, 1961, 1969a, b; Boucot and others, 1964; Moench, 1970a, b, 1971, 1973; Ludman, 1976; (foot-note continued on the next page)

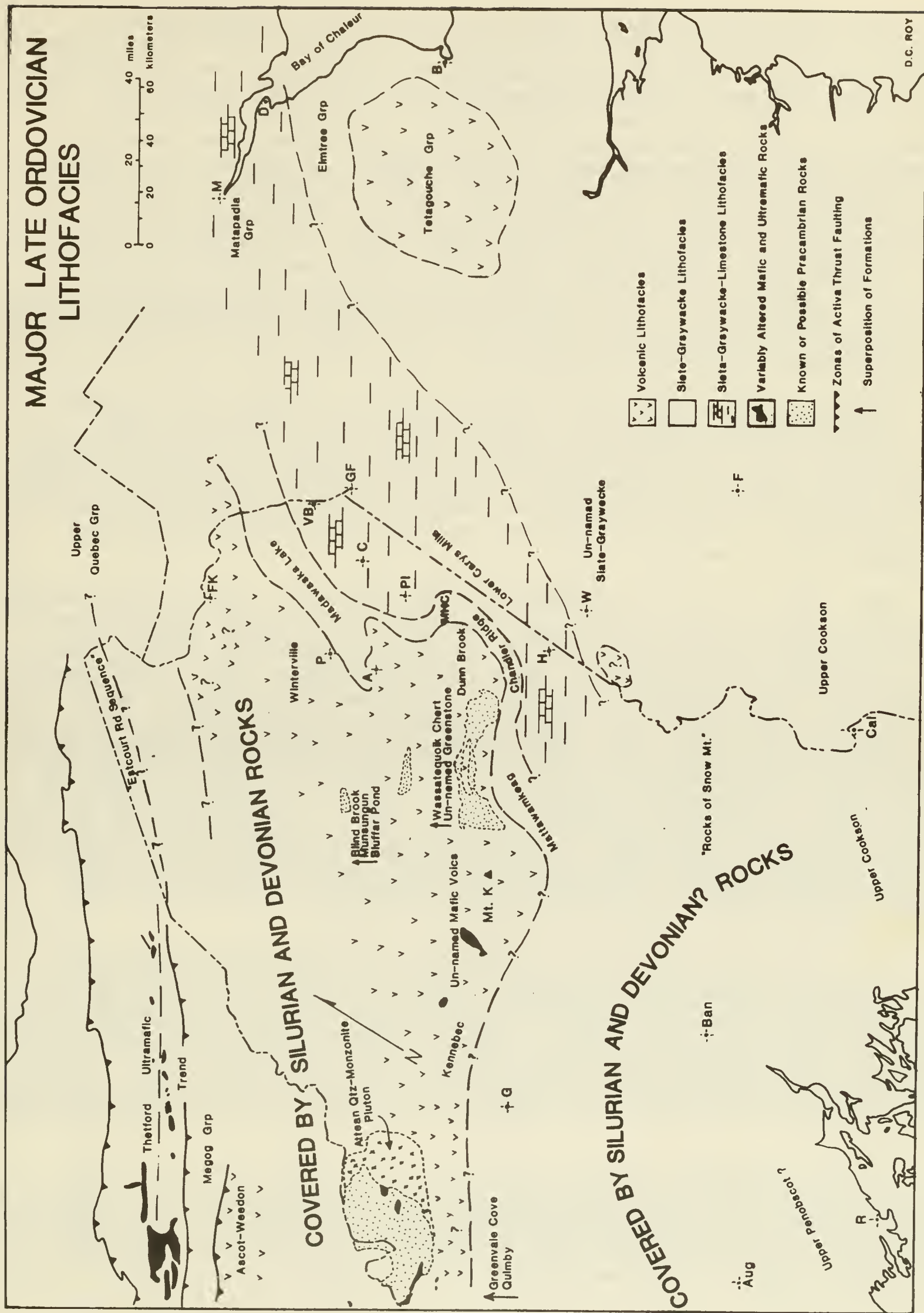


Figure 3: Distribution of major Late Ordovician (Late Caradocian largely) lithofacies in Maine and adjacent Canada prior to the Taconian Orogeny.

The volcanic lithofacies forms a broad tract in northwestern Maine. The western limit of the lithofacies in much of northwestern Maine is obscured by Siluro-Devonian rocks but it is possible that the lithofacies is continuous in the subsurface with the Ascot-Weedon volcanic sequence in Quebec. In northern-most Maine the volcanic lithofacies is represented by the Winterville Formation (Roy, Trip B-6, this volume) but the lithofacies is embodied in a variety of named and un-named units to the southwest as indicated in Figure 3. The volcanic lithofacies in Maine consists largely of fragmental and pillowed spilitic basalts, dacite, karatophyre, soda rhyolite and apparently less abundant interstratified black sulfidic slate, chert, and graywacke. Hynes (1976) assigns the volcanic rocks of the Winterville and Bluffer Pond formations to the alkali olivine basalt suite based on titanium concentrations in pyroxene. Hynes has found the un-named volcanic rocks of the Weeksboro-Lunksoos Lake Anticline to be more varied in silica content and to be transitional between alkaline and non-alkaline rocks.

Variably altered Mafic and ultramafic rocks are found in the volcanic belt southwest of the region of the conference. As discussed by Boudette and Boone (1976) some of these rocks, especially those marginal to the Chain Lakes Massif, may be part of a dismembered pre-Middle Ordovician ophiolite sequence; others appear to be younger intrusions. Massive sulfide deposits of the Kuroko type are present locally within the volcanic lithofacies but have not been described in detail as yet.

East of the Volcanic Lithofacies is a belt of more or less contemporaneous slate and graywacke. The slate-graywacke lithofacies is found in the Madawaska Lake Formation in the north (Figure 3) but is also seen in the Chandler Ridge (Pavlides, 1968) and Mattawamkeag (Ekren and Frischknecht, 1967) formations farther south. The Mars Hill Conglomerate (Pavlides, 1978) may be a part of this lithofacies located beneath the limestone-rich Carys Mills Formation of the Aroostook-Matapedia Belt. The Quimby-Greenvale Cove sequence of western Maine is considered here to be expressions of this lithofacies. The graywacke beds of this belt contain abundant volcanic detritus which presumably is derived from volcanic islands in the western volcanic belt. In the vicinity of Portage, Maine, the Winterville Formation interfinger's with the Madawaska Lake Formation as discussed by Roy (Trip B-6, this volume). No evidence of penecomtemporaneous deformation (e.g. subduction zone tectonism) is evident along the lithofacies boundary at Portage or elsewhere in Maine.

East of the slate-graywacke lithofacies and forming much of the anticlinal terrain of the Aroostook-Matapedia belt is the thick limestone-rich Carys Mills Formation. The lower part of this formation is known to represent the Upper Ordovician (Pavlides, 1968) and thus it forms the next lithofacies. In the Presque Isle-Caribou area Roy (1978; Trip B-6, this volume) has found the lower part of the Carys Mills to be more graywacke-rich than the upper part. Both the slate and graywacke of this lower phase are calcareous and are interlayered with

¹Ludman and Griffin, 1974; Pankiowskyj and others, 1976; Osberg, 1968; Gates, 1961; Ruitenberg and Ludman, 1978; Noble, 1976; Lee and Noble, 1977; Greiner, 1973; Greiner and Potter, 1966; Rast and Stringer, 1974; Ayrton, and others, 1969; Skinner, 1964; Naylor and Boucot, 1965; Pavlides and others, 1968; Lajoie and others, 1968.

minor micritic limestone. The lower Carys Mills forms the slate-graywacke-limestone lithofacies of Figure 3. The lithofacies may extent into central Maine but it can only be inferred to extend a short distance south of Houlton based on the distribution of the Carys Mills Formation as a whole.

East of the Carys Mills belt, largely in New Brunswick, rocks of Late Ordovician age are generally assigned to the Tetagouche Group. Most of the subdivisions of the Tetagouche Group as originally described by Helmstaedt (1971) for the Bathurst-Newcastle area have been extended into southwestern New Brunswick by Venugopal (1979) and Rast, Lutes, and St. Peter (this volume). In the Miramichi zone near Woodstock, N.B., slate and graywacke units of the Tetagouche Group appear to comprise the section above the quartzite-rich sequence that forms the basal unit of the group. In so far as these slate-graywacke units are Late Ordovician in age, they may represent an analogous lithofacies to that found west of the Aroostook-Matapedia belt. Presumably the eastern slate-graywacke lithofacies interfingers with the mafic volcanics of the "type" Tetagouche in the Bathurst area. At present, however, it is difficult to be certain of lithofacies relationships east of the Aroostook-Matapedia belt because of major faults along the northwestern margin of the Miramichi Anticlinorium (Rast and Stringer, 1974) and the paucity of fossils within the anticlinorium.

The Taconian Orogeny

The Taconian Orogeny in northeastern Maine was a relatively mild deformational event but it produced substantial changes in paleogeography and sedimentation in the Aroostook-Matapedia basin. The region of the conference is southeast of the Thetford Ultramafic trend (Figure 3) that marks the suture zone of the Taconian Orogeny as envisioned by St. Julien and Hubert (1975). The southeastern extent of Taconian deformation in Maine appears to coincide roughly with the transition from the volcanic to the slate-graywacke lithofacies. In the eastern townships of Quebec the Taconian began in the Early Ordovician and proceeded through the Late Ordovician with the development of large-scale overthrust tectonics apparently associated with a complicated obduction/subduction history. In northern Maine the Taconian was less dramatic and is seen to be younger, namely latest Ordovician and earliest Silurian (Roy, Trip B-6, this volume).

No metamorphic fabric of Taconian origin has been reported in northern Maine from any of the lithofacies belts. Cleavage in Ordovician slate is essentially axial planar to Acadian folds. Volcanic rocks and graywacke beds of Ordovician age (and younger for that matter) are generally incleaved north of the latitude of Presque Isle but show increasing foliation along strike to the south. The cleavage in these more competent rocks is also consistent with Acadian deformation.

Richter and Roy (1974, 1976) argue that Taconian metamorphism of the Winterville Formation probably did not exceed prehnite-pumpellyite grade. Assessments of Taconian metamorphism elsewhere in northern Maine have not been made and become more difficult as Acadian metamorphic grade increases southwestward.

The rocks of the Aroostook-Matapedia basin have produced no clear evidence

of Taconian deformation. Early recumbent folds reported by Rast, Lutes, and St. Peter (this volume) in the Carys Mills near Woodstock represent pre-Acadian deformation that may be as old as Taconian. However, it is also possible that these early folds are non-tectonic and represent syndepositional slump folds as described elsewhere in the Carys Mills (see Stringer and Pickerill, this volume) or they may be tectonic folds associated with Salinic deformation at the close of the Silurian (see below).

Taconian deformation of the Miramichi zone is well established (Rast and Stringer, 1974; Rast and others, this volume) and probably carried the Tetagouche rocks to at least the greenschist grade locally (Helmstaedt, 1971). Skinner (1974) has proposed an early northwest-trending fold system followed by the generation of younger more open folds with northeast-trending axes for the Bathurst area but he does not attempt to assign ages to the fold events.

Lower Silurian Lithofacies

Following the Taconian Orogeny substantial paleogeographic changes are revealed by Llandoveryan lithofacies. Figure 4 shows the distributions of the major lithofacies for the Late Llandoveryan-Early Wenlockian time period and the approximate limits of emergent areas. A paleogeographic map for the Early Llandoveryan produced by Ayrton and others (1969) is similar to that shown for a slightly later time period in Figure 4. One major difference between the early and late Llandoveryan lithofacies pictures is the presence of the upper "ribbon-rock" facies of the Carys Mills in the interior of the Aroostook-Matapedia basin during the earlier part of the Llandovery. The Late Llandoveryan-Wenlockian interval was selected for Figure 4 because of the larger amount of paleontologic data available and greater complexity of the lithofacies pattern.

The land area called "Taconia" was probably contiguous with the pre-Taconian North American Craton since the ocean-basin closure described by St. Julien and Hubert (1975) appears to have completely removed marine conditions from northwestern Maine and adjacent Quebec. The thick sequence of coarse clastic sediments of the Frenchville-Rangeley belt represents deposits along the newly established continental margin in Maine. In northeastern Maine these coarse clastic sequences give way eastward to much finer grained deep-water lithofacies in the Aroostook-Matapedia Basin (Roy, Trip B-6 and C-5, this volume). Similar off-shore lithofacies changes can be seen from western to central Maine along the western flank of the Merrimack Synclinorium.

The Miramichi terrain appears to have also been uplifted by Taconian deformation as evidenced by coarse sediments deposited in the Bathurst Basin (Noble, 1976). The extent of "Miramichia" in southwestern New Brunswick is not well known and faulting has been largely responsible for limiting our understanding of the detail lithofacies picture in the Woodstock area.

It is of interest to note that by Late Llandoveryan time the broad Merrimack Synclinorium in Central Maine seems to have been divided by Miramichia into two basins, the Aroostook-Matapedia Basin and the Fredericton Basin, in New Brunswick as pointed out by McKerrow and Ziegler (1971) who figured a map similar to that shown in Figure 4. McKerrow and Ziegler (1971) suggest that the Fredericton Basin was the main oceanic continuation of the Merrimack Basin and that the Aroostook-Matapedia Basin was a branch that terminated in the Gaspé Peninsula.

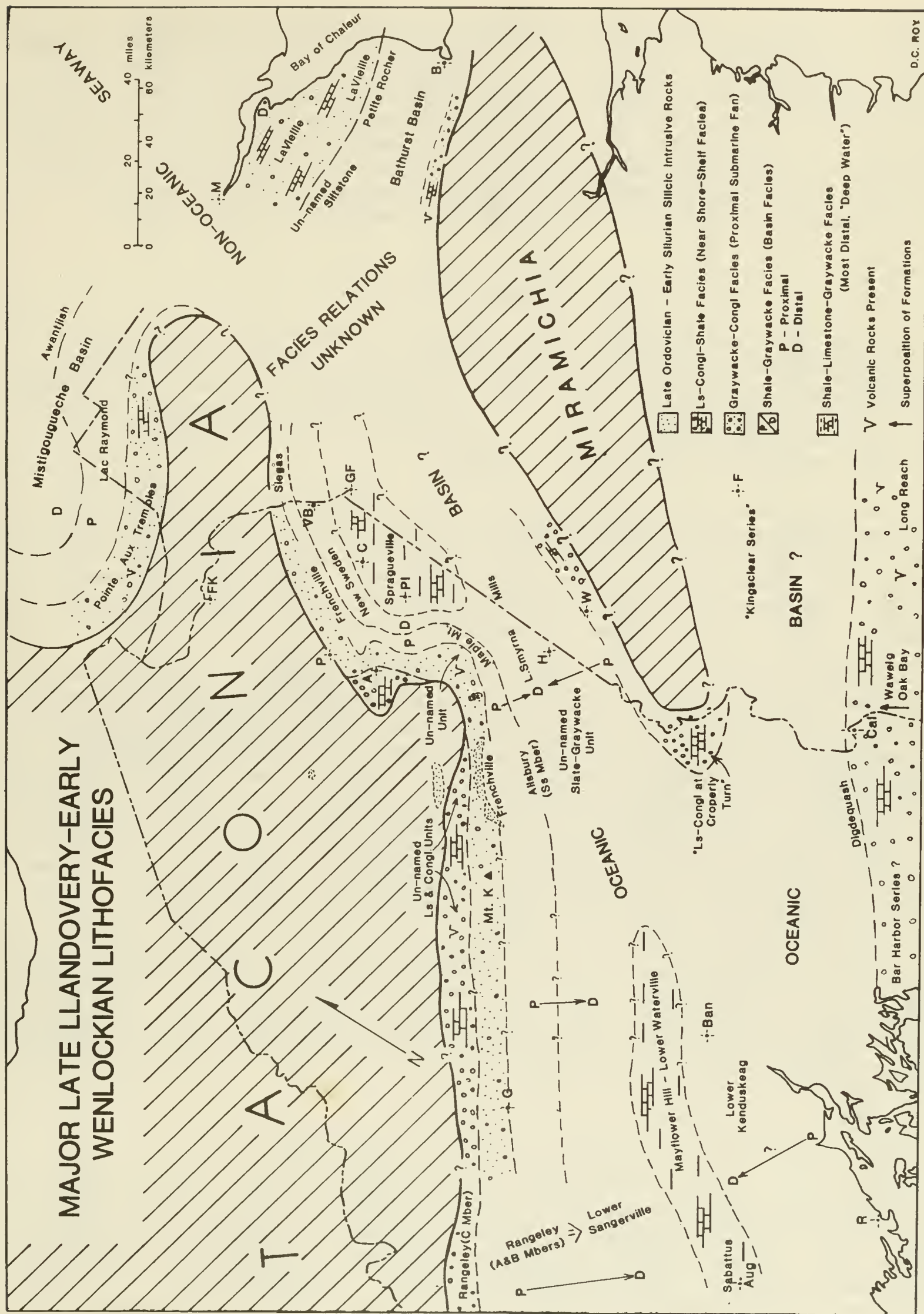


Figure 4: Distribution of major Late Llandovery-Early Wenlockian lithofacies in Maine and adjacent Canada following the Taconian Orogeny.

As shown in Figure 4, the relationships of the deep-water facies of the Aroostook-Matapedia Basin in Maine to more shallow-water sedimentary sequences in the Bay of Chaleur are unclear. The northeastern Maine basinal Silurian is similar to the sequence in the Temisconta-Matapedia region of Quebec as described by Lajoie and others (1968) that are here shown as deposited in the "Mistigougueche Basin". The Silurian of the Mistigougueche Basin was probably everywhere deposited on Taconian-deformed pre-Silurian rocks unlike the basinal sequences in Maine. It is possible that the Silurian Aroostook-Matapedia sequence in Maine is continuous with the Bathurst Basin as described by Noble (1976).

Late Silurian Lithofacies

By the Late Wenlockian as shown in Figure 5, elevation of the Taconian land areas had been greatly reduced by erosion and possibly by subsidence. Everywhere in the deep-water basins that were established in the Early Silurian the deposits of Late Silurian age indicate more distal environments. In addition, shallow-water deposits appear to be widespread in the previously upland areas, especially during the Ludlovian.

Small islands, including Somerset Island of Boucot (1961, 1969), were probably common on the inundated Taconia as Late Silurian transgression took place. These islands provided the coarse detritus for sandstones and conglomerates that are common in the shallow-water sequences. Intermediate and mafic volcanic rocks interlayered with the shallow-water sediments are also common and suggest at least some tectonism during the advance of marine conditions.

Eastward in the basin one finds a considerable reduction in conglomerate deposition and the development of generally thin-bedded flysch typified by the Jemtland, Upper Smyrna Mills, Upper Allsbury, and Upper Sangerville formations. The sandstone beds within the basins are predominantly calcareous, rusty-weathering quartzofeldspathic graywacke. The graywacke beds become increasingly more lithic and abundant toward the west. The Late Silurian sandstones on the west side of the basin, extending from Augusta to beyond Van Buren (Figure 5), are clearly derived from the somewhat subdued Taconia to the west. Derivation of similar sandstones in the southeastern part of the Merrimack Basin from an eastern "platform" may have begun in the Early Silurian and continued into the Ludlovian (Figure 5; McKerrow and Ziegler, 1971; Ludman and Griffin, 1974; Ruitenberg and Ludman, 1978). The two-sided character of the Aroostook-Matapedia Basin is less well documented for the Late Silurian than for the Early Silurian (Figure 4) but is probable. Derivation of Late Silurian detritus from Miramichia in the Bay of Chaleur area is well known (Greiner and Potter, 1966; Greiner, 1973; Noble, 1976; Lee and Noble, 1977) and Anderson (1968) reports clasts of Miramichia slate and graywacke in Silurian conglomerates northeast of Woodstock, New Brunswick.

The Salinic Disturbance and Earliest Devonian Lithofacies

The Salinic Disturbance (Boucot, 1962) is indicated by a disconformity in the Presque Isle area between the Jemtland Formation and Early Devonian (Late Gedinian) volcanic and sedimentary rocks of the Dockendorf Group (Boucot and others, 1964). Unconformities between shallow-water Late Silurian (Ludlovian) sedimentary rocks and quite similar rocks of Late Gedinian age are common also along the Munsungun-Pennington Mountain Anticlinorium (Hall, 1970; Boone, 1970;

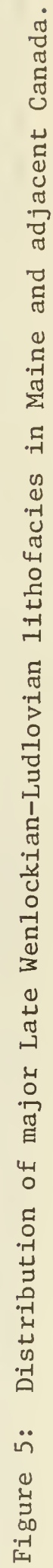




Figure 6: Distribution of major Late Gedinnian lithofacies in Maine and adjacent Canada following the Salinic Disturbance.

Roy and Mencher, 1976). Between these two regions where unconformities between Silurian and Devonian rocks are present there is a belt in which Devonian rocks apparently rest conformably on the Silurian. In the Ashland Synclinorium the Fogelin Hill Formation which probably contains the systemic boundary rest conformably on the Jemtland (Roy and Mencher, 1976). Hall (1970) suggests that the Siluro-Devonian boundary lies within the Third Lake Formation in the southeastern part of the Spider Lake region. It is therefore possible that an elongate basin in northeastern Maine survived the Salinic uplifts which caused erosion both to the northwest and southeast of the basin as shown in Figure 6..

The extents of the syn-Salinic basin are not well established. To the northeast the basin may correspond to the lower Cape Bon Ami Formation which seems to rest conformably on the St. Leon Formation (Lajoie and others, 1968). To the southwest it is not possible to trace the "axis" of the basin very precisely or even to be sure of its persistence beyond the Third Lake area. The axis is presumed to have passed northwest of the position of Mount Katahdin during Predolian-Early Gedinian time. By Late Gedinian (Figure 6) the basin appears to have widened considerably. It is generally presumed that the slate-rich sequence in central Maine extends up into the Devonian. By the Late Gedinian (and possibly earlier) a connection between the northern basin in which the Fogelin Hill was deposited and a basin in central Maine is possible. The dimensions of the central Maine basin are uncertain and depend on the true extent of Early Devonian rocks in central and eastern Maine as discussed below.

Much of central and southern New Brunswick was probably uplifted during the Salinic interval (Figure 6). This conclusion is based on the apparent absence of sedimentary rocks of Early Devonian age in that region and the presumption that the "Kingsclear Series" of the Fredericton Basin does not extend into the Devonian. The present writer believes that this emergent area extended well into southeastern Maine because it is suspected that the Devonian is absent there also. Boucot (1968) has essentially come to the same conclusion. Unfossiliferous formations such as the Flume Ridge, Bucksport, Fall Brook, and Brighton are lithologically more similar to dated Silurian units farther north and northwest than they are with dated Devonian sections. In addition, the early recumbent folds postulated by Osberg (this volume) may be Salinic folds in a completely Silurian section that were later refolded by more upright Acadian folds. If the units just mentioned are indeed restricted to the Silurian, the syn-Salinic basin may have been restricted to west-central Maine where the Madrid, Solon, and similar units are present.

Whatever the extent of the eastern land area, an important belt of volcanic and sedimentary rocks developed along northwestern edge in New Brunswick and northeastern Maine. The Dalhousie Group of the Bay of Chaleur region and the Dockendorf Group of the Presque Isle area lie along a more or less continuous and broad volcanic belt that is immediately post-Salinic (Boucot and others, 1964). Naylor (this volume) suggests that this belt represents a continental volcanic arc. The Dockendorf Group has been shown to be about 3600 meters thick (Boucot and others, 1964) and the Dalhousie section is at least 600 meters thick and possibly thicker. Substantial subsidence must have accompanied the formation of these thick sections of volcanic rocks and interlayered shallow-water sedimentary rocks.

Rapid westward facies changes are present in the Late Gedinian rocks

between Presque Isle and Ashland. In Ashland mudstone conglomerate, limestone, polymictic conglomerate and sandstone beds are interlayered with slate. The mudstone conglomerates are usually monomictic with clasts of either volcanic rocks or clasts from the Jemtland Formation (some containing graptolites).

In the Matagamon Lake area (see Neuman and Rankin, this volume) there is good evidence of a large delta involving the Seboomook Formation and Matagamon Sandstone (Hall and others, 1976). Hall and others (1976) conclude that the deltaic sediments were derived from the east beginning in the Gedinian, and that the delta prograded westward during the Siegenian. The delta implies a major upland to the east and the presence of a relatively deep basin offshore to the northwest. Additional such deltas with coarsening-upward sections may be present in western Maine (e.g. the Madrid-Carrabassett sequence of Boone, 1973). Such delta sequences may reflect continuation of the southeastern source area into the Maine Slate Belt.

Siegenian Time and The Seboomook Formation

The term "Seboomook" has been applied over the years to the widespread gray slate and graywacke sequences in northwestern Maine. Boucot (1970) has reviewed the distribution of such sequences in Maine and elsewhere within the northern Appalachians and summarized the paleontological control on their ages.

In the Ashland-Portage area and in much of northwestern Maine, the Devonian section is divisible into two parts. The lower part is a lithologically variable section made up of conglomerate, lithic sandstone and limestone interlayered with cleaved mudstone or slate. This lower part usually is fossiliferous (both fauna and flora) and up to about 1 km or so thick. The upper part consists of generally fine-grained, well-cleaved slate with minor but typically cyclically interlayered thin beds of graywacke. The upper part is unfortunately rarely fossiliferous and has no defined "top".

The Seboomook of the Matagamon Lake Area (Hall and others, 1970) and the Moose River Synclorium (Boucot, 1961, 1969), on the other hand, coarsens upward and is succeeded by fossiliferous shallow-water sandstone units (Matagamon and Tarratine sandstones) followed by silicic volcanic sequences (Traveller Rhyolite and Tomhegan Formation). These stratigraphically "topped" Seboomook sections are difficult to reconcile with the Seboomook further north and west that seems to not have a definable stratigraphic top and indeed may have been "terminated" by the Acadian Orogeny. As pointed out by Boucot (1970), very careful mapping within this monotonous sequence must be done in order to subdivide the now broad "Seboomookland" into subbasins and sources of sediment supply.

The work of Hall and others (1970) is just such an effort. At the time of his death, Ely Mencher was also attacking this very problem in a broad subdivision of "Seboomookland" west of Ashland. Ely's approach was to try and piece together the facies variations in the lower, better-dated part of the section in a large region where fossils were likely to be found due to low regional grade. Ely's field style and meticulous attention to detail were well suited to the task and it is a shame he could not finish it.

The Acadian Orogeny

The principal fold and cleavage producing event in northern Maine post-dates Siegenian deposition and is assigned to the Acadian Orogeny. The Mapleton Formation of Late Middle Devonian age (Schopf, 1964; White, 1975; Boucot and others, 1964; see Naylor, this volume) rests unconformably on nearly vertical lower Devonian and Silurian rocks and provides the best stratigraphic evidence for the end of the orogenic phase of the Acadian.

Folds

The style of Acadian folding varies from place to place depending primarily on the rocks involved in the folded sections (Pavlides and others, 1964; Roy, 1970; Pavlides, 1974). Where thick volcanic or sandstone-conglomerate sections are present the folds are more open and concentric and flexure-slip mechanisms predominated. The Chapman Syncline (involving the Dockendorf Group), the Stockholm Mountain Syncline (involving the Frenchville Formation), and the Castle Hill Anticline (involving both the Winterville and Frenchville formations) are good examples of major folds that are simple in form (see Figures 1 and 2, Roy, Trip B-6, this volume). Where the stratigraphic section is dominated by pelite more tightly appressed, generally symmetric and steeply plunging folds of either concentric or similar form are present. Pelite mobility in these folds is indicated by common pelite injection along cleavage planes in limestone and graywacke beds as well as hinge thickening of pelite beds (Pavlides, 1965; Roy, 1970).

Disharmonic folding is seen locally (Pavlides, 1973) and is common on an outcrop scale in the laminated Silurian ironstones (Roy, 1970). Disharmonic folding on a large scale has been suggested for the Chapman Syncline (less competent pre-Devonian versus the competent Devonian volcanic rocks) by Pavlides (1974) and may be characteristic of the Pennington Mountain Anticlinorium where a thick previously deformed and volcanic-rich Winterville Formation is overlain by the thick slate-rich Seboomook Formation.

Pavlides (1965, 1971, 1972) has documented complicated fold geometries (including overturned fold plunges) in the Ordovician and Silurian pelite-rich units in the Aroostook-Matapedia Anticlinorium. Outcrop size and spacing have so far precluded complete analysis of possible pre-Acadian folding in these rocks, but they are generally thought to have undergone only one deformation (Pavlides, 1974, Roy, 1970). Early folds may be more common than previously recognized, but reports are rare (Hamilton-Smith, 1970; Rast, Lutes and St. Peter, this volume) and their distinction from non-tectonic folds is difficult.

Cleavage

A prominent S_1 cleavage is present in the pelitic rocks. The S_1 cleavage is generally parallel or nearly parallel to axial planes of Acadian folds. Massive sandstone and conglomerate beds or volcanic rocks in the northern part of the conference area typically show no cleavage; thin graywacke beds in slate-rich sections, however, do show a fracture cleavage. Southward from the latitude of Presque Isle sandstone beds and volcanic rocks show increasingly well developed foliation.

Cleavage in pelitic intervals varies from fracture cleavage in which there is a low degree of micaceous alignment to flow cleavage in which there is pervasive orientation of micaceous minerals parallel to cleavage surfaces. In subgreenschist slates fracture cleavage is characteristic of non-calcareous slate whereas flow cleavage is typical of calcareous slate. Calcareous slates are usually phyllitic in appearance. Pressure solution effects, as discussed by Stringer and Pickerill (this volume), are probably of widespread importance in cleavage formation, especially in low-grade slates.

A late northwest trending fracture cleavage (S_2) is associated with the broad Houlton Oroflex which folds the axial surfaces of the Acadian folds in the vicinity of Houlton (Pavrides, 1974).

Metamorphism

Acadian metamorphism in the northern part of the region of the conference did not exceed prehnite-pumpellyite grade (Combs and others, 1970; Richter and Roy, 1976). The grade there increases from the prehnite-analcime subfacies in the northwest (Pennington Mountain Anticlinorium) to the pumpellyite-epidote-Actinolite subfacies to the south and southeast. Metamorphism continuous to increase southward from Presque Isle so that in the Bridgewater-Houlton area low greenschist grade is present (Pavrides, 1965, 1971). Similarly, to the south and southwest, grade becomes low greenschist in the Spider Lake area Hall, 1970) and in the Weeksboro-Lunksoos Lake Anticline (Neuman, 1967).

The prehnite-pumpellyite paragenesis in the north, which is seen only in volcanic rocks and in lithic graywacke beds, appears to be prograde. It is not clearly associable with an Acadian fabric since the rocks containing the diagnostic minerals are not foliated. A regional increase in fluid CO_2 content appears responsible for suppression of the diagnostic minerals of the facies in volcanic rocks of suitable composition in the Presque Isle area (Richter and Roy, 1976). Confirming the southeastward temperature increases suggested by the changes in diagnostic sub-greenschist minerals are preliminary paleotemperature results by Anita Harris of the USGS (Personal communication, 1976) on Silurian and Devonian conodonts. Her results suggest a temperature increase from $50^\circ C$ (lower Devonian Square Lake Limestone) in the prehnite-analcime zone to $190-250^\circ C$ within the pumpellyite-epidote-actinolite zone in the Castle Hill Anticline and Chapman Syncline (Lower Silurian Frenchville and Spragueville formations).

Plutonism

Widely scattered granitic Acadian plutons are present in the area covered by the conference. The two largest plutons are the Katahdin Quartz Monzonite (Hon, this volume; Neuman and Rankin, this volume) and the Bottle Lake Complex (Ayuso and Wones, this volume); both will be examined during the conference. North of these large plutons are several smaller intrusives: the Nickerson Lake and Pleasant Lake plutons south of Houlton (Pavrides, 1971); the Munson Pluton just northwest of Presque Isle (Boucot and others, 1964); the Deboulli Stock northwest of Portage (Boone, 1962); and the Chandler Lake Pluton southwest of Ashland.

All of the plutons cut the Acadian cleavage and have well-defined metamorphic aureoles. For the most part the plutonic rocks are not strongly foliated.

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OUTLINE OF THE PLEISTOCENE GEOLOGY OF NORTHERN MAINE
AND ADJACENT CANADA

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INTRODUCTION

For some time, questions concerning the intensity, style, and limits of glaciation throughout northern New England, adjacent Canada, and the Maritime Provinces have been the subject of much controversy.

Initially, Chalmers (1895) put forth the concept of glaciation based on the influence of local centers of ice accumulation. These local ice centers were to have coalesced at maximum glaciation but were thought to have still remained partially independent. As investigations continued this view remained viable and is still considered tenable (Prest and Grant, 1969; Grant, 1977) although modified with respect to specifics concerning limited isostatic rebound, areas showing limited or no traces of glaciation, and stratigraphy.

Alternatively, because of characteristics and features related to strong regional ice flow, lithologic transport, and alpine glaciation, Goldthwaite (1924) advanced the concept of continental glaciation which was controlled by southward flowing lobes of ice emanating from a massive continental ice sheet.

This latter view has been expanded to a more sophisticated level by using modern day analogues such as Antarctic (Hughes, 1973; Borns, 1979, pers comm.) and Greenland (TenBrink, 1974) glaciers, along with concepts of calving bays (Thomas, 1977), ice streams (Hughes et al., 1977) and, more recently, models employing the concept of the thermal regime of ice masses (Sugden, 1977; Hughes, 1979, pers. comm.).

Currently, both perspectives are receiving renewed stimulus.

With the exception of the classic work of Leavitt and Perkins (1935) northern Maine has only recently begun to be mapped on a reconnaissance basis, primarily because in the past, access was difficult in what were generally regarded as "wilderness" areas. However, land use planning considerations in conjunction with accelerated logging operations led to various programs starting several years ago involving reconnaissance surficial mapping (Prescott, 1973a; 1973b). It was primarily through these programs, fostered and sponsored by the Main Geological Survey, that northern Maine became an active area of Pleistocene geological investigation. Fortunately so, for prior to these programs most Pleistocene investigations were concentrated along southern and coastal Maine leaving a large, unmapped void between that area and Canada, where Canadian colleagues were working.

As it turns out, this region of northern Maine and adjacent Canada seems destined to play a pivotal role in the deciphering of Pleistocene events along the northeastern corridor of the United States and Canada.

Areally, northern Maine consists of the largest counties in the state; Aroostook County (the largest) forms the northwest, north, and eastern borders of Maine as it abuts Quebec and New Brunswick, separated from Canada to the north by the St. John River (international boundary); Franklin and Somerset Counties delimit the western border of Maine where they join the Quebec border which is roughly coincident with the Boundary Mountains; Piscataquis and Penobscott Counties constitute the central - southern region of northern Maine.

For the most part, the region is underlain by the cyclically bedded grey slate and metasandstone Seboomook Formation. Some large areas of metamorphosed volcanic rocks occur scattered throughout the region. In central Piscataquis County biotite - muscovite granites and quartz monzonites outcrop, increasing in occurrence to the south. Metamorphosed sandstone, siltstones, and limestones outcrop near the eastern Maine-New Brunswick border (See Figures 1 and 2 of Roy, Trip B-6, this volume)

Topographically, the area is a dissected upland plateau with regional bedrock structures trending northeasterly to southwesterly, transverse to known ice-flow direction.

STRATIGRAPHY AND EVENTS

The stratigraphy of glacial deposits of southern Quebec and the events which they represent (Table 1) is based on work by McDonald and Shilts (1971). They determined that the last glacial advance in southern Quebec is represented by the Lennoxville Till, having a northwest provenance, and that southeastern Quebec was deglaciated by about 12,500 years BP..

This Lennoxville Till is correlated with the surface till of Maine (Borns and Calkin, 1977). Coastal Maine was covered by ice 13,200 years BP. (Schlee and Pratt, 1970) and readvances to the coast occurred during general ice recession (Borns, 1966; 1973). An active ice margin existed along the eastern coast of Maine (Borns, 1966) approximately 13,300 years BP.. At about the same time an active ice margin existed in southwestern New Brunswick (Gadd, 1973). Both McDonald and Shilts, and Borns and Calkin consider the Lennoxville Till as representing the entire late Wisconsinan time interval. Borns and Calkin (1977) suggest that the final deglaciation of Maine occurred by thinning and stagnation of ice throughout the uplands of northwestern Maine with no reorganization of flow to form an active center of ice dispersal. In support of this contention, they suggest that cirque glaciers did not develop in the mountains of west-central Maine following dissipation of the ice. This implies that the regional snow line had become elevated above the highest mountains prior to their emergence. This view is supported by Davis (1976) but is in opposition to views held by Thompson (1960, 1961a, 1961b) and Caldwell (1966).

Borns and Calkin (1977) visualize that by the time the ice margin had retreated to the proximal side of the coastal moraine belt all

Time-Stratigraphic Unit		Rock-Stratigraphic Unit	Chronologic Control
Wisconsin Stage	late	post-Lennoxville Sediments	12,640 \pm 190 - peat (GSC-312) 12,570 \pm 220 - peat (GSC-419) 12,000 \pm 230 - marine shells (GSC-936) 11,500 \pm 160 - marine shells (GSC-475-2)
		Lennoxville Till	
	middle	Gayhurst Formation	> 20,000 B.P. (GSC 1137) Ca 4000 varves
		Chaudière Till	
	early	Massawippi Formation	> 54,000 B.P. (Y-1683) > 41,000 B.P. (GSC-507) > 40,000 B.P. (GSC-1084)
		Johnville Till	
		pre-Johnville Sediments	

Table I. Quaternary Stratigraphic Column, Southeastern Quebec; (after McDonald and Shilts, 1971)

Canadian ice flow was southward within the active ice mass north of the Boundary Mountains. During this period of stagnant ice in Maine, the Canadian ice constructed successive frontal moraines as the ice margin receded to the St. Lawrence lowland of Quebec (Gadd, 1964; McDonald and Shilts, 1971).

Glacial Lake Madawaska was formed in New Brunswick when the St. John River at Grand Falls, New Brunswick was dammed by the Grand Falls Moraine (Lee, 1955). Recession of the ice front from Grand Falls had to have occurred well before 10,200 years BP. because peat overlying till at Green River, New Brunswick has been dated at $10,200 \pm 350$ years BP. (Lee, 1955). Kite (1979) has had this peat redated and concludes that it possibly could be as old as approximately 12,000 years BP. Until recently (Genes and Newman, 1979; Brewer, 1980) end moraines had not been reported between the end moraine belt around Pineo Ridge, at coastal Maine, and the international border at the west side of the Boundary Mountains. Therefore, the limitations imposed by previous investigations of adjacent areas required the conclusion that all of the surface till of Maine be Lennoxville in age, that is, it is a simple remnant of the last major advance to the coast. In turn, this implied that subsequent to the invasion of the Champlain Sea, approximately 13,000 years BP., Laurentide ice did not transgress into Maine (Lasalle et al., 1977) or even at all during late Wisconsinan time (Grant, 1977).

It must be mentioned that there are extremely divergent views as to the behaviour of the ice in southeastern Canada. Gadd (1973) envisions the Highland Front Moraine system in Quebec as having been emplaced as a recessional moraine during normal retreat of the margin of southerly flowing ice. Lasalle et al. (1977) visualize the system as having formed from northerly flowing ice which resulted from the division of the Laurentide ice sheet into separate lobes by the Champlain Sea embayment. Grant (1977) because of his adherence to the concept of Appalachian ice centers maintains that Laurentide ice never reached Maine.

Field mapping in northern Maine and southern Canada now implies that deglaciation was more complex than previous work has suggested.

Evidence from striations, moraines, outwash overrun and capped by till, and three or possibly four distinct tills can be interpreted as having resulted from multiple glaciations; from different flow regimes within a single ice sheet; or from penecontemporaneous deposition from different ice centers.

The following considerations are implicit in any interpretation regarding the late Wisconsinan history of northern Maine and adjacent Canada:

- 1) A late Wisconsinan readvance formed the Pineo Ridge Moraine along the eastern coast of Maine at approximately 12,700 years BP. (Borns, 1966). The minimum age of deglaciation of southwestern New Brunswick is implied to be $12,600 \pm 279$ years BP. (Gadd, 1973).
- 2) The date of 12,600 years BP. for the Highland Front Moraine (Gadd, 1964) in Quebec, represents the minimum date for the incursion of the Champlain Sea into the St. Lawrence Valley. This date also represents the minimum date at which the Laurentide ice became detached from the Laurentide ice north of the St. Lawrence River.
- 3) All observed evidence indicates that northern Maine was overrun by ice moving in a southeasterly direction, and that glacial recession was accomplished by downwasting and recession from coastal to northern Maine. Evidence in northwestern Maine indicates ice flow to the north.

In light of these controversies, this years' surficial field trips should prove interesting indeed.

Claude Gauthier has been deciphering the New Brunswick Pleistocene for several years and will lead a trip on the New Brunswick side of the St. John River Valley. Glacial stratigraphy around Edmunston, a

reappraisal of Glacial Lake Madawaska, and sections of the Grand Falls Moraine complex, will be discussed. (Trip B-9)

Kite and Borns will discuss deposits associated with Glacial Lake Madawaska as they relate to both sides of the middle reaches of the St. John River. Although, primarily concerned with Holocene events, they suggest important concepts regarding the determination of deposits and how they should be mapped. (Trip C-4)

Wisconsinan stratigraphy along the southern bank of the St. John River, including multiple till sections which permit interpretation of late Wisconsinan deglaciation in northern Maine, is the subject of the trip by Genes and Newman. (Trip B-8)

Genes, Newman, and Brewer will examine moraines, eskers, and other landforms in northern Maine which suggest the mode and extent of till emplacement in that region. (Trip C-6)

D. Caldwell will review the Wisconsinan alpine glaciation of Mt. Katahdin. In case a climb is not possible because of inclement weather, a surficial trip of the area is planned. (Trips A-3, B-2)

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TRIP A-1

GEOLOGY OF THE BOTTLE LAKE COMPLEX, MAINE

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Introduction

The Bottle Lake Complex consists of two late Paleozoic, massive granitic plutons exposed in an area of about 1100 km² in the Scraggly Lake, Springfield, Winn, Wabassus Lake, Nicatous Lake, and Saponac 15-minute quadrangles, and located between U.S. Rts. 6 and 9 in east-central Maine (Figure 1). The Complex comprises two overlapping, subcircular, generally east-west trending bodies intruding greenschist facies metamorphic rocks of the Merrimack Synclinorium. The bodies have a granitic extension striking to the northeast (Figures 1, 2).

Larrabee and others (1965) mapped the granitic and metamorphic rocks in the Big Lake region in reconnaissance. Their study of the granitoid rocks represented by the Bottle Lake Complex made no field or petrographic distinctions between these rocks except for the possibility of an internal contact between the main body of the Complex and the large northeast-trending granitic protuberance (Topsfield facies). Observed field and petrographic differences were ascribed to secondary processes, mainly as a result of country-rock assimilation. However, Ayuso (1979) recognized several mappable petrographic types that exhibited a consistent arrangement and indicated the presence of two separate intrusive granitoid plutons. From east to west, the two granitoid plutons consist of the granite of Whitney Cove and the granite of Passadumkeag River, respectively. The Topsfield facies is assigned to the Whitney Cove pluton because of similarity in their petrographic and field relations and absence of internal contacts (Figure 2).

Metamorphic rocks in the region are presently under intensive scrutiny by A. Ludman (1978a, 1978b) and coworkers in the area to the northeast and east of the Complex, and by Wones (1979) in the Norumbega fault region to the south. Detailed mapping by Olson (1972) to the southwest of the Complex and the reconnaissance work of Doyle and others (1961) and Cole (1961) concentrated respectively, to the southwest and north. As summarized by Ludman (1978b), metamorphic rocks in this area are in the greenschist facies and exhibit variable age and lithology (Figure 2). These rocks range in age from Cambro-Ordovician (?) to Siluro-Devonian, and from almost monomineralic sandstones to andesitic volcanics. Most traverses from country-rock toward the granite show well-developed mineral zonation as a result of the contact metamorphic effects (Ayuso, in press).

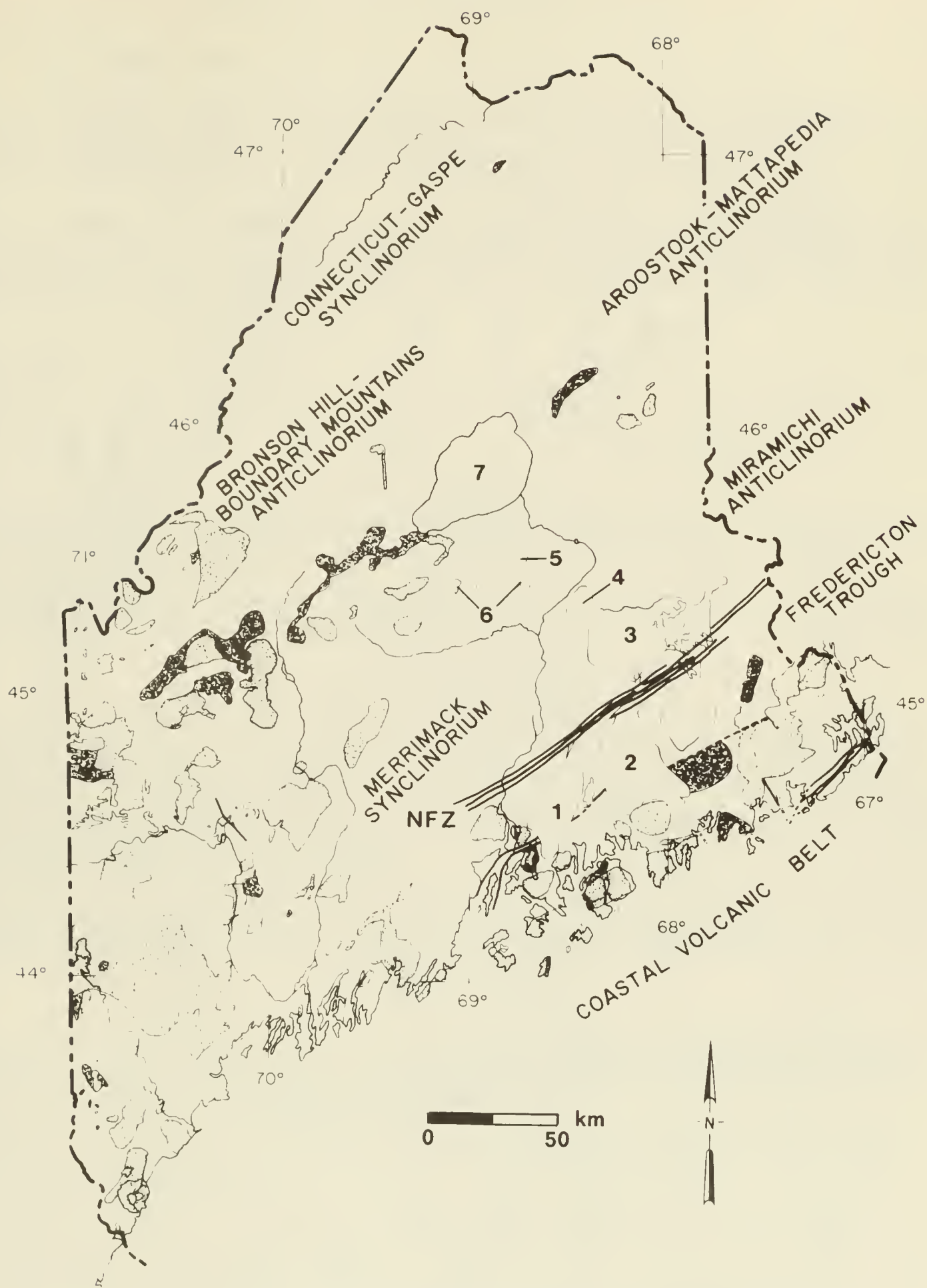


Figure 1 - State of Maine showing the location of the Bottle Lake Complex. Light stipple - granite and granodiorite; dark stipple - diorite and gabbro. 1-Lucerne pluton; 2-Lead Mountain pluton; 3-Bottle Lake Complex; 4-Center Pond pluton; 5-Seboeis foliated granitoid; 6-Seboeis non-foliated granites; 7-Katahdin pluton. NFZ-Norumbega Fault Zone. (from Loiselle and Ayuso, 1980).

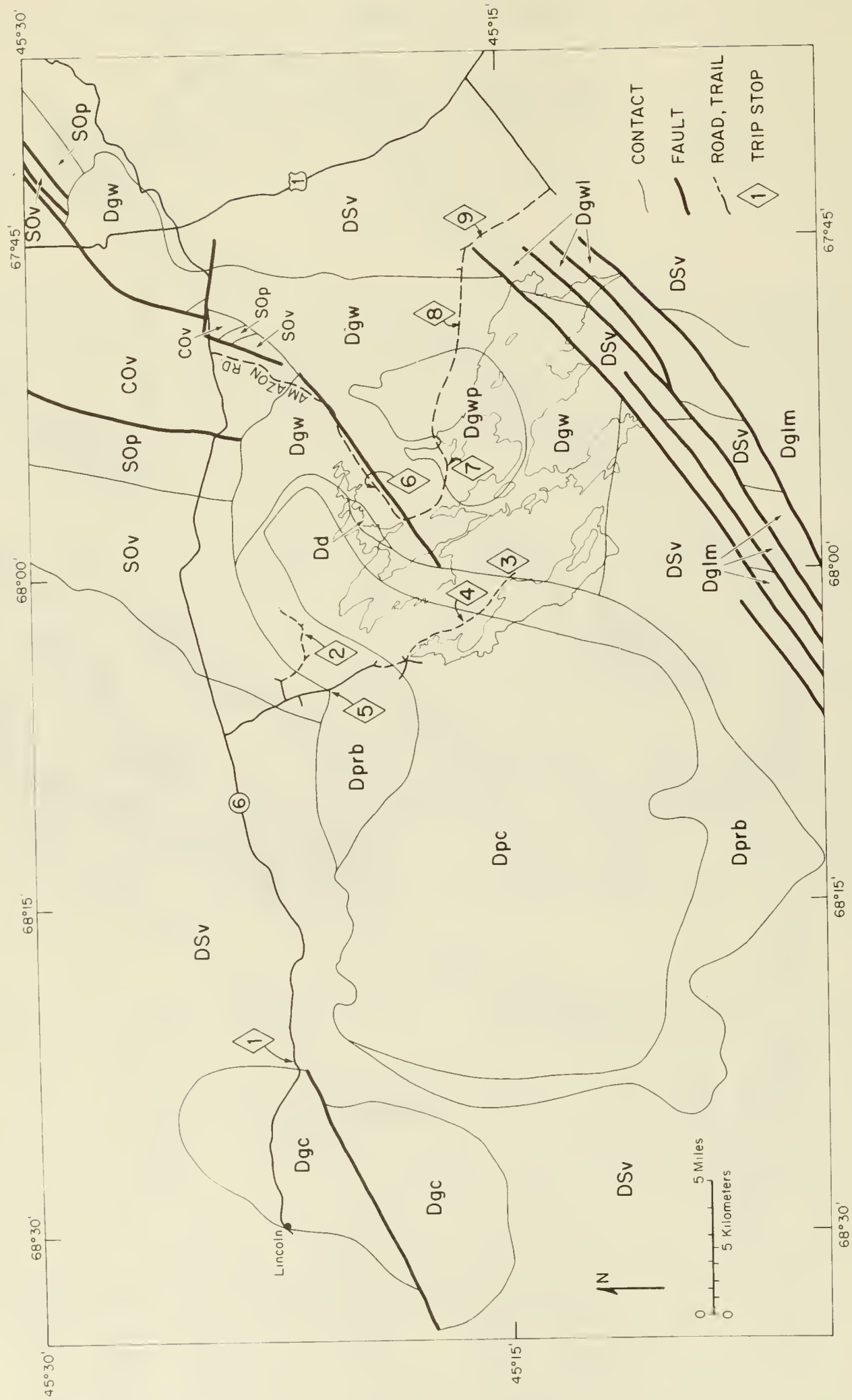


Figure 2 - Generalized geologic map of the Bottle Lake Complex. Field trip stops are also shown.

PLUTONIC ROCKS - Bottle Lake Complex

Granite of Passadumkeag River

Dpc

Coarse-grained biotite-hornblende bearing facies - generally quartz monzonitic composition, high color index, euhedral amphibole prisms, and abundant mafic inclusions

Dprb

Coarse-grained biotite-hornblende granitic to quartz-monzonitic facies - commonly heterogeneous in texture; quartz forms large mosaic textures, generally poorer in mafic minerals than rocks in the interior

Amphibolite

Dd

Diabase (?) intruded in cataclastic zone

Granite of Whitney Cove

Dgwp

Medium to coarse-grained, biotite bearing facies - homogeneous in composition, typically porphyritic, and low color index

Dgw

Coarse-grained biotite facies granite - equidimensional texture, low color index, biotite in pseudo-hexagonal books

PLUTONIC ROCKS - Other Granitic Rocks

Dgc

Granite of Center Pond - medium-grained, granitic rock ranging from diorite to biotite granite; contains abundant hornblende

Dglm

Undifferentiated granitic rocks, usually coarse-grained, biotite-hornblende bearing rocks

Dwl

Medium-grained granite found only within Norumbega fault zone, lower mafic-mineral content than Bottle Lake Complex

METAMORPHIC ROCKS

DSv

Vassalboro Fm (?) - calcareous siltstones and pelites showing prominent mineralogic zonation from contact metamorphic effects imposed by the Bottle Lake Complex

SOv

Undifferentiated Siluro-Ordovician (?), volcanic rocks, rusty pelitic beds

SOp

Undifferentiated Siluro-Ordovician (?), pelitic siltstones containing graded beds

COu

Graywakes, siltstones, slates are the main lithologies

Geology by R. A. Ayuso and assistants (1977, 1978, 1979); based on fieldwork by D. Larrabee and assistants (1965), A. Ludman (1978a,b), D. R. Wones (1979), and Scambos (1980).

Field Relations

Field and petrographic contrasts between the Whitney Cove and Passadumkeag River plutons include their strikingly different textures and mineralogy, particularly the kind and abundance of ferromagnesian phases, abundance of aplites and pegmatites, and abundance and type of inclusions. Field relations are summarized in Ayuso (in press). With the exception of amphibole, both granites are generally coarse-grained, consist of two feldspars, biotite and quartz, with minor amounts of primary sphene, zircon, allanite, apatite, oxides (magnetite and ilmenite) and sulfides. Sharp internal contacts, dikes and inclusions of older in younger granite are notably absent, but the relative sequence of intrusion may be established from mapping petrographic types, cross-cutting relationships of faults, and preliminary age determinations.

The granite of Whitney Cove

The granite of Whitney Cove covers an area of about 400 km². Its mineralogy is consistently granitic (Streich, 1973), has relatively low total ferromagnesian mineral content, lacks amphibole and shows either a predominant seriate (rim facies) or porphyritic (core facies) texture. Aplites and pegmatites are generally common in most exposures, while mafic inclusions are smaller and less common than in the Passadumkeag River pluton.

This pluton consists of three units: 1) Topsfield facies, 2) rim facies, and 3) core facies. The generalized geologic map shown on Figure 2 does not separate the Topsfield facies from the rim facies of the Whitney Cove pluton because of their similarity in field relations and absence of an internal contact between them. The Topsfield facies is commonly a medium to coarse-grained, reddish rock of low color index, approximately 400 m.y. old as determined by U-Pb work on zircons (R. Zartman, pers. comm.). Away from the main body of the pluton of Whitney Cove, it becomes finer-grained, progressively pinker, lower in color index and more intensely sheared. The rim facies is a pink, medium to coarse-grained, hypidiomorphic, seriate rock of low color index. No mineralogical characteristics distinguish the core from the rim, but textural differences are, however, significant and may be used to delineate the extent of each; contacts between the two are for the most part gradational. Within the core rocks, biotite is sometimes present as phenocrystic clots, but it is more abundant as fine-grained aggregates disseminated through the rock. The predominant phenocrysts are alkali-feldspars (up to 3.5 cm) and these are usually euhedral but also show edges embayed by biotite, plagioclase and quartz; some of the alkali-feldspars show rapakivi textures. Plagioclase mantled by alkali-feldspar (anti-rapakivi texture) is rarely present in these rocks. Quartz is usually constrained to the matrix, however, it is also present as a phenocryst (up to 1 cm) phase. Biotite is characteristically enclosed within subhedral, phenocrystic (up to 2.5 cm) plagioclase; less commonly, biotite forms euhedral phenocrysts similar to those found in the rim facies. Fine-grained biotite and plagioclase make numerous clusters in the matrix and give this rock an easily recognizable speckled appearance.

The consistently granitic composition of the Whitney Cove pluton is shown in Figure 3, where the modal variation obtained by counting stained slabs is displayed. The quartz-plagioclase-alkali feldspar ternary shows complete overlap between core and rim facies, and this is reinforced in the accompanying ternaries between mafic and felsic minerals. Also note the low scatter in total ferromagnesian content and the absence of a clear concentration gradient of mafic-poor to mafic-rich rocks from rim to core (Figures 4,5). Both facies have approximately 4% total mafic mineral content. The general mineralogical homogeneity of the pluton is further shown by its lack of correlation between alkali-feldspar and mafic-mineral content in Figure 5.

The granite of Passadumkeag River

The two facies in the pluton of Passadumkeag River are the rim and core (Figure 2). Generally pink rocks of seriate to equidimensional texture, finer-grained and lower in color index characterize the rim. Amphibole is not abundant but it is usually present, at least in trace quantities. The low content of ferromagnesian phases and many of the textures are reminiscent of the rim facies of the Whitney Cove granite. Preliminary Rb-Sr isotopic determinations on whole-rock samples suggests an age of about 360 ± 16 m.y.

In contrast, rocks found in the interior of the pluton are invariably mafic-rich, commonly showing abundant, euhedral amphibole, and numerous mafic-schist, metasedimentary, and quartz-dioritic inclusions. Characteristically, this facies consists of gray porphyritic rocks with prominent subhedral to euhedral alkali feldspar phenocrysts (3 to 5 cm) that are usually embayed and mantled by plagioclase in a typical rapakivi texture. Many of these phenocrysts contain inclusions of euhedral plagioclase, quartz, and ferromagnesian minerals which help to delineate several growth zones. Euhedral phenocrysts of plagioclase (3 to 5 cm) are commonly filled with inclusions; together with fine-grained ferromagnesian minerals, plagioclase forms clots that alternate with quartz-rich clusters in the matrix. A common characteristic of the Passadumkeag River granitoid is that euhedral amphibole (0.5 cm) commonly encloses biotite.

In contrast to the Whitney Cove pluton, marked variations exist between the rim and core facies of the granite of Passadumkeag River (Figure 3). In this case, a granitic rim envelops a quartz monzonitic core of consistently higher plagioclase and mafic-mineral content. As shown in Figure 4, the core facies has almost double the mafic-mineral content of the rim; this zonation is clearly exhibited in the reversed correlation of alkali-feldspar and mafic content between rim and core (Figure 5).

The Whitney Cove pluton generally contains more felsic dikes (aplites, pegmatites, granophyres) than the granite of Passadumkeag River. These dikes are typically less than 0.7 cm wide, exhibit diffuse contacts, and are concentrated near the granite-country rock and granite-granite contact.

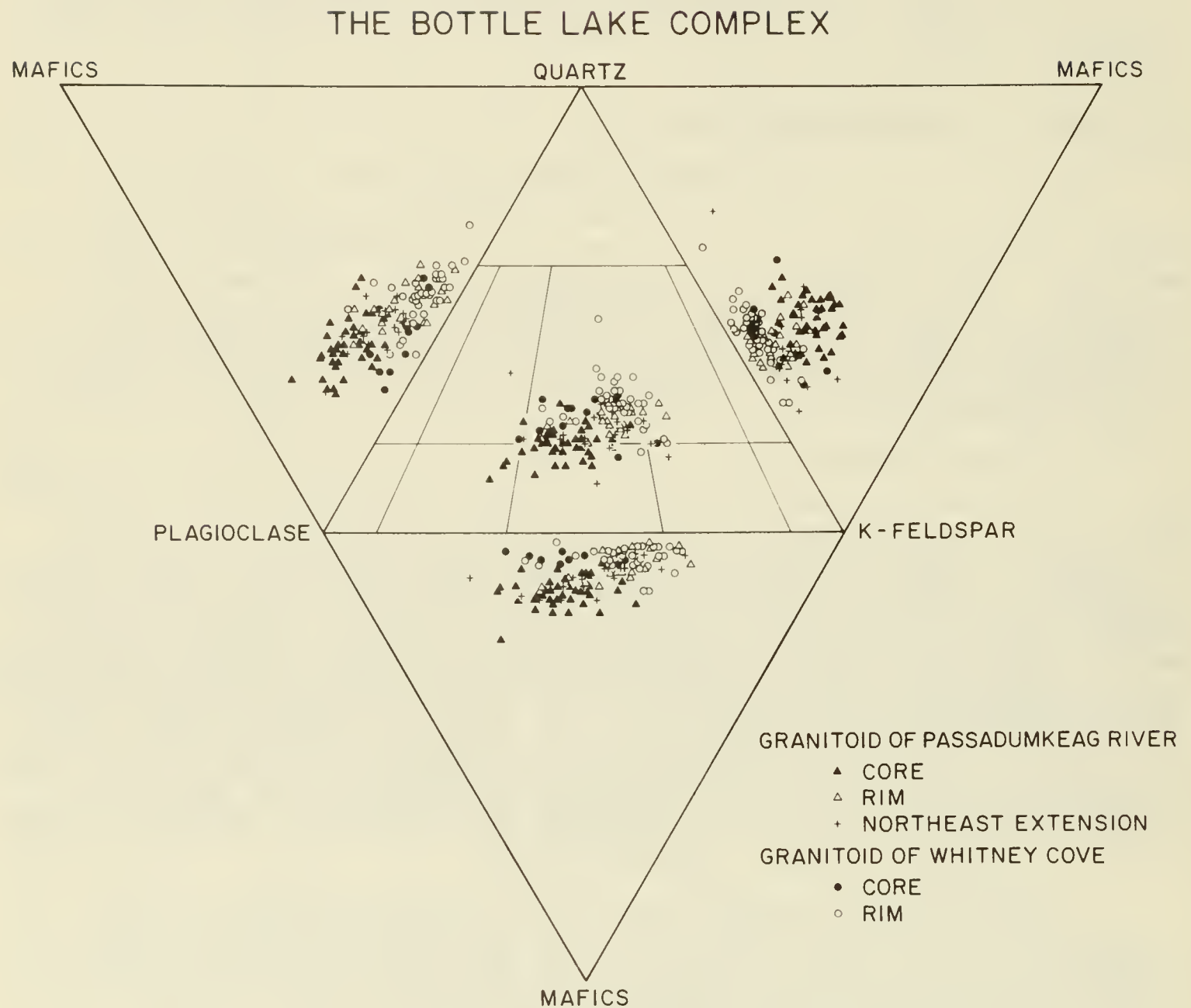


Figure 3 - Modal mineralogy of the Bottle Lake Complex. Northeast extension refers to the granitic rocks of the Passadumkeag River pluton intruded north of the cataclastic zone.

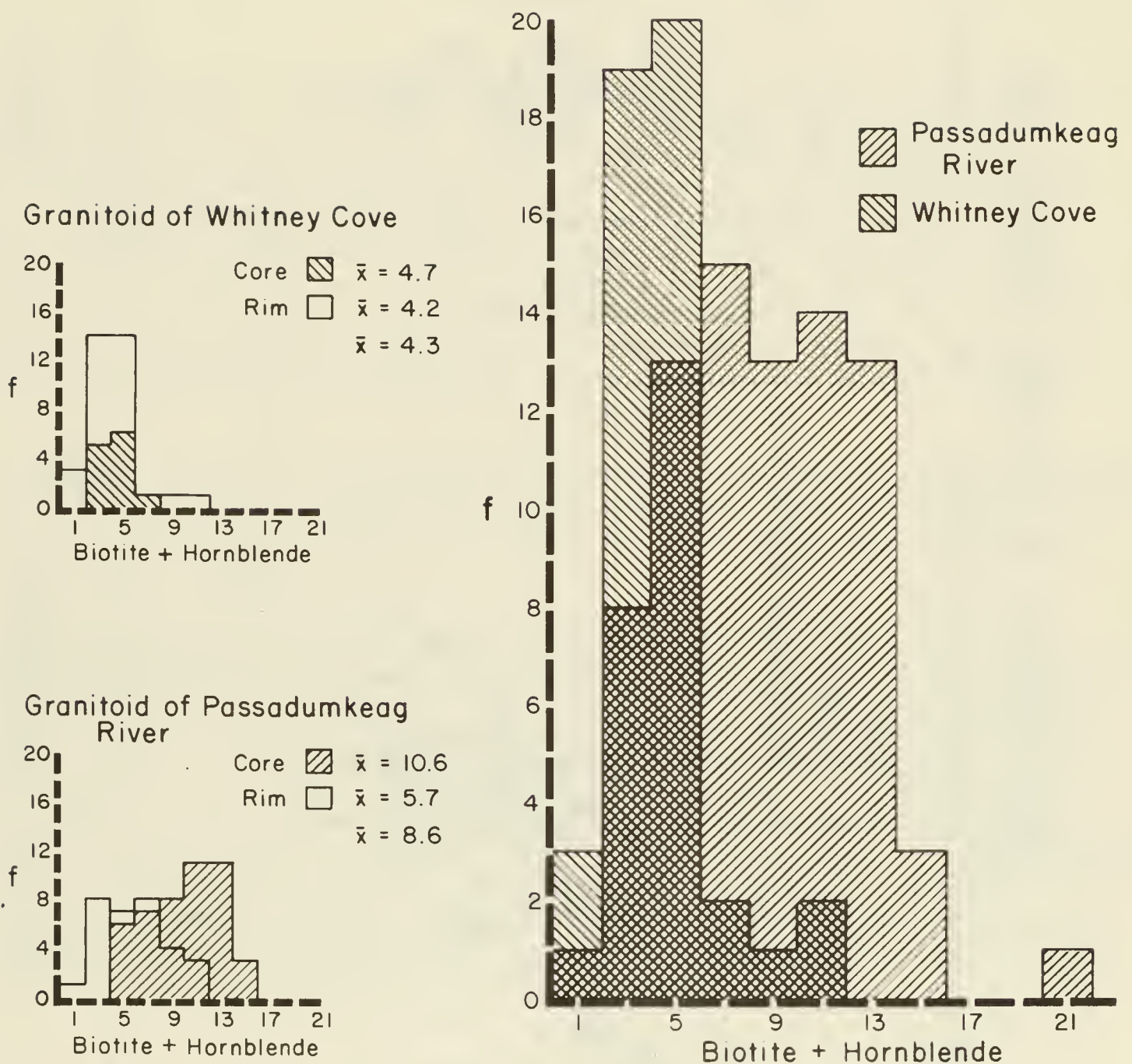


Figure 4 - Histogram of the total abundance of mafic minerals in the Bottle Lake Complex

THE BOTTLE LAKE COMPLEX

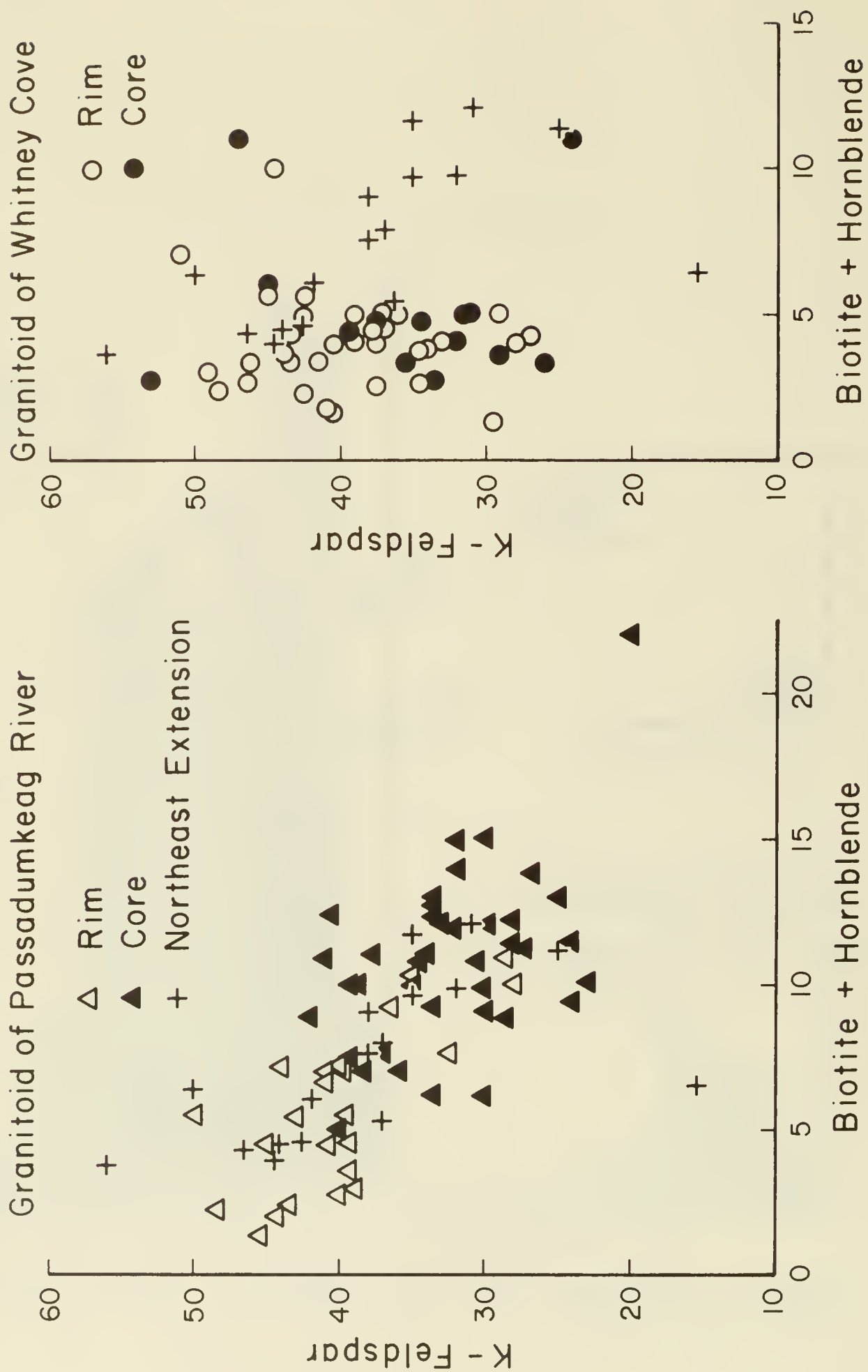


Figure 5 - Plot of alkali-feldspar vs. mafic mineral content of the Bottle Lake Complex. Northeast extension refers to the granitic rocks of the Passadumkeag River pluton intruded north of the cataclastic zone.

An amphibolite unit of limited length outcrops between the two plutons (Figure 2), and it is concentrated near the fault zone that cuts the Whitney Cove pluton. Typical rocks associated with this unit range from black to green in color and contain prominent orange feldspar phenocrysts; these rocks are intimately mixed with granitic rocks of the Whitney Cove pluton. Aplites, pegmatitic pods and large inclusions of variable lithology (from felsic to mafic mineralogy) are commonly found within the amphibolite unit.

Structures

The Bottle Lake Complex is generally massive in fabric, often even at the granite-country rock contact. Several areas within the Complex show well-developed foliation. These areas may be grouped as follows:

1) randomly distributed zones in the core facies of the Passadumkeag River pluton; 2) the area associated with the amphibolite rocks near a zone of cataclastic deformation; and 3) the randomly foliated rocks in the core of the Whitney Cove pluton. The c dimensions in the feldspars are aligned parallel to the northeast in the core rocks of the Passadumkeag River granite and in the amphibolite rocks. In contrast to this, no discernible preferred orientation in the regional sense is apparent in the foliated rocks found in the core of the Whitney Cove pluton.

Although joints and fractures are common in the Complex, they typically show great scatter even within individual outcrops. The only exception to this lack of preferred alignment is found in those joint sets associated with cataclastic zones.

Three cataclastic zones are present in the Bottle Lake Complex:

1) the Norumbega fault zone, 2) an unnamed EW fault, south of the Topsfield facies and exposed near U.S. Rt. 6, and 3) the NE-trending belt that cuts the Whitney Cove pluton (Figure 3). The Norumbega fault system forms the southern contact of the Whitney Cove granite. According to Wones (1979), the amount of displacement along the fault is unknown. The EW fault is exposed near U.S. Rt. 6 between the main mass of the Whitney Cove granite and the Topsfield facies and shows left-lateral displacement (Ludman, 1978b). Exposure of this fault is expressed by sheared, epidotized, red rocks typical of the Topsfield facies. Finally, the NE-trending cataclastic zone through the pluton of Whitney Cove forms a wide band (1.5 km?) which terminates against the Passadumkeag River granite.

Bulk Chemistry

Preliminary major and trace element X-ray fluorescence analyses of the Bottle Lake Complex are shown in Figure 6. In spite of the significant mineralogical contrasts between the two plutons, the data suggests a re-

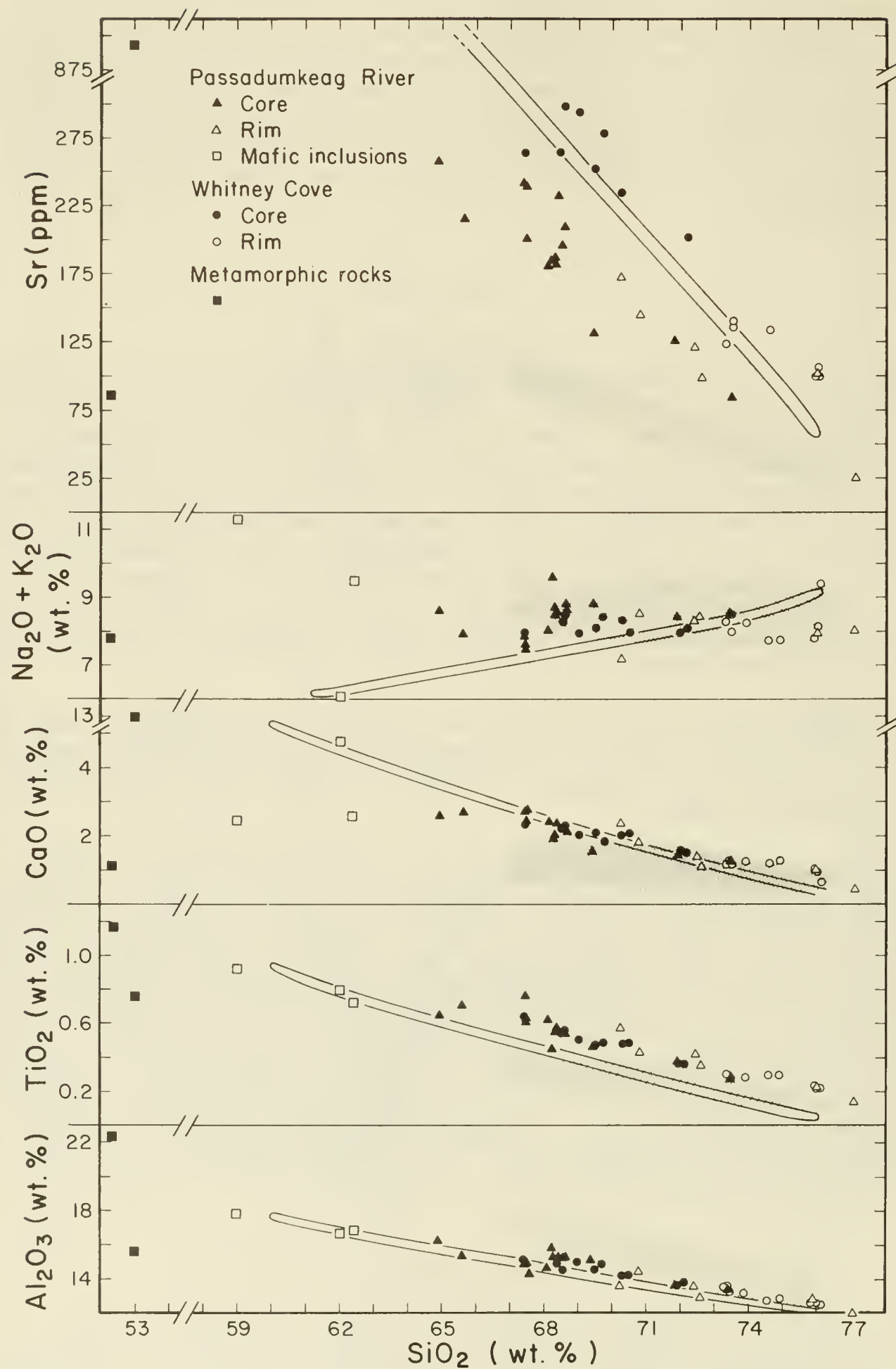


Figure 6 - Silica variation diagrams of selected oxides and strontium of the Bottle Lake Complex. Stippled field represents analyses of the Lucerne and Lead Mountain granites; field outline represents the Center Pond pluton (Scambos, 1980).

markable colinearity in the trend of both granitoids, showing similar ranges in SiO_2 content, and complete overlap between data for samples from the two cores and rims. As expected, the core of the Passadumkeag River granite is less silica-rich and more mafic in character than its rim; the pluton shows reverse zonation. However, the Whitney Cove core facies, despite its mineralogical similarity to the rim, is extremely unlike it in composition, and instead resembles the core of the Passadumkeag River pluton. Reverse zonation is chemically expressed in both plutons, but it is mineralogically evident only in the Passadumkeag River granitoid.

Strontium concentrations reinforce the contrast between core and rim of these plutons (Figure 6). As in the major element trends, a reverse zonation is clear as cores are substantially enriched in strontium compared to the rims. Unlike the major element plots, however, a difference is suggested in the core of the Whitney Cove pluton which consistently shows higher strontium content compared to the Passadumkeag River core. An additional suggestion from this data is, in spite of the overlap, that each pluton describes parallel but distinct evolution trends from strontium-rich to strontium-poor rocks.

Mineral Chemistry

Biotite is the most common ferromagnesian phase in the Bottle Lake Complex; in spite of the different petrographic character among the rocks of the Complex, no consistent differences are expressed in its optical properties. However, on the basis of microprobe analyses of representative samples from each granitoid, biotite shows distinct compositional variations and trends. In general, biotite in the Whitney Cove pluton is lower in $\text{Fe}/(\text{Fe} + \text{Mg})$ ratio (0.55 to 0.65) than that in the Passadumkeag River granite (0.60 to 0.80). Also, biotite from Whitney Cove tends to have higher F, but lower TiO_2 . Figure 7 shows that both plutons have decreasing Ti with increasing Al^{VI} and overlap between cores and rims. Biotite in the Passadumkeag River pluton shows a larger range in Al^{VI} compared to that in Whitney Cove; near the granite-country rock contact, it tends toward extreme enrichment in Al^{VI} .

Amphibole is concentrated within the Passadumkeag River pluton and generally shows a lack of correlation in traverses from the edge of the core toward the interior. As in biotite, amphibole shows a limited range in $\text{Fe}/(\text{Fe} + \text{Mg})$ from 0.55 to 0.75 without a distinct range within each facies. Figure 8 shows that amphibole compositions tend to increase in Al^{IV} with increasing A site content. A prevailing feature of the Passadumkeag River pluton is the common intergrowth and inclusion of biotite in amphibole. According to the studies of Wones and Dodge (1977) on potassic magmas, crystallization of biotite prior to amphibole is facilitated by water deficient conditions.

Preliminary work on feldspar compositions show that plagioclase ranges from pure albite to An_{50} for the Complex as a whole. Most plagioclase cores are usually richer in anorthite, show normal zonation, and are enveloped by strongly zoned albitic rims. Between the two plutons, the Whitney

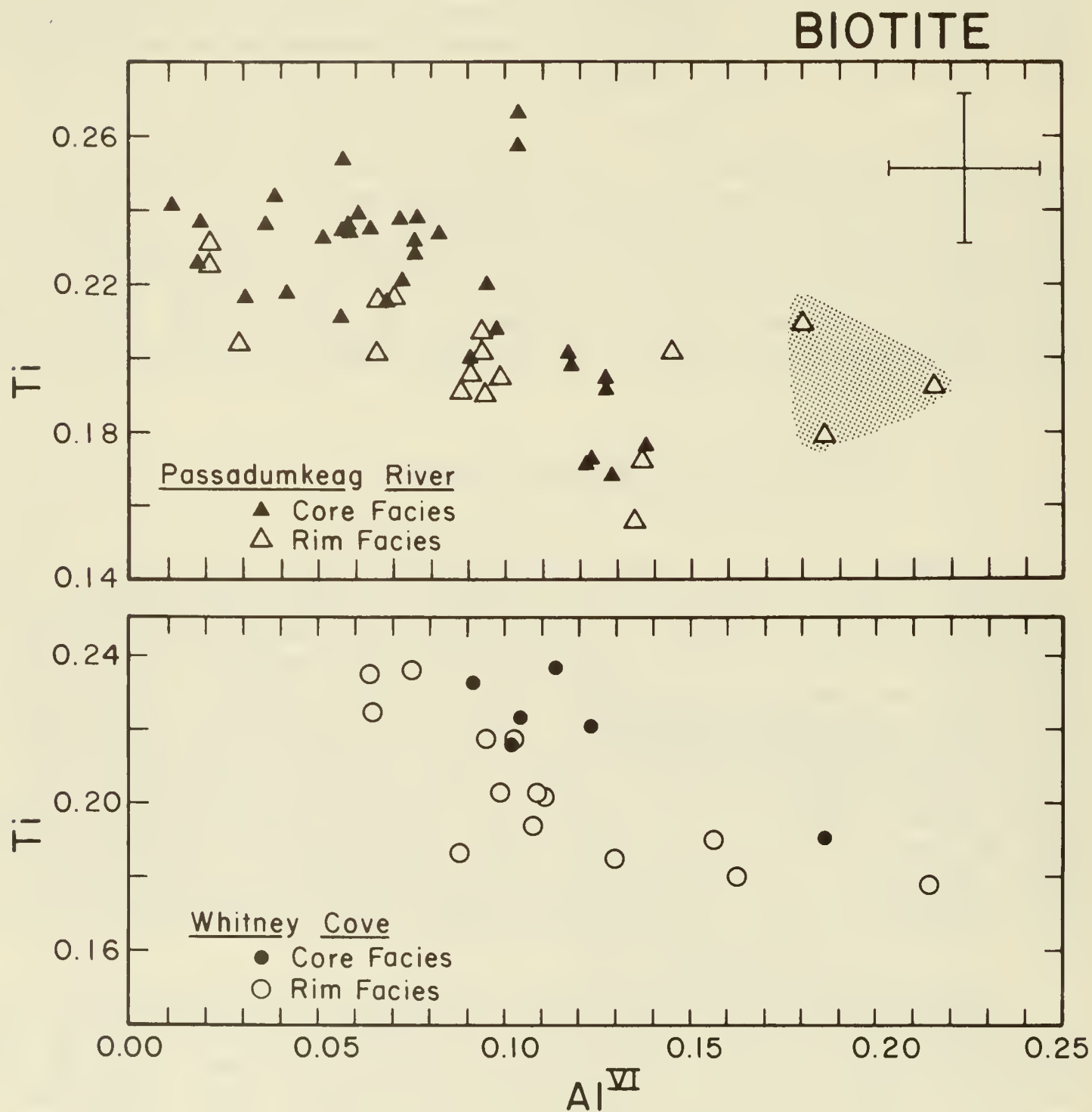


Figure 7 - Representative microprobe analyses of biotite from the Bottle Lake Complex showing cation proportions of Ti and Al^{VI} in one-half of the formula unit. The stippled field represents biotite near the granite-country rock contact. Error in determinations shown approximately at 2σ level.

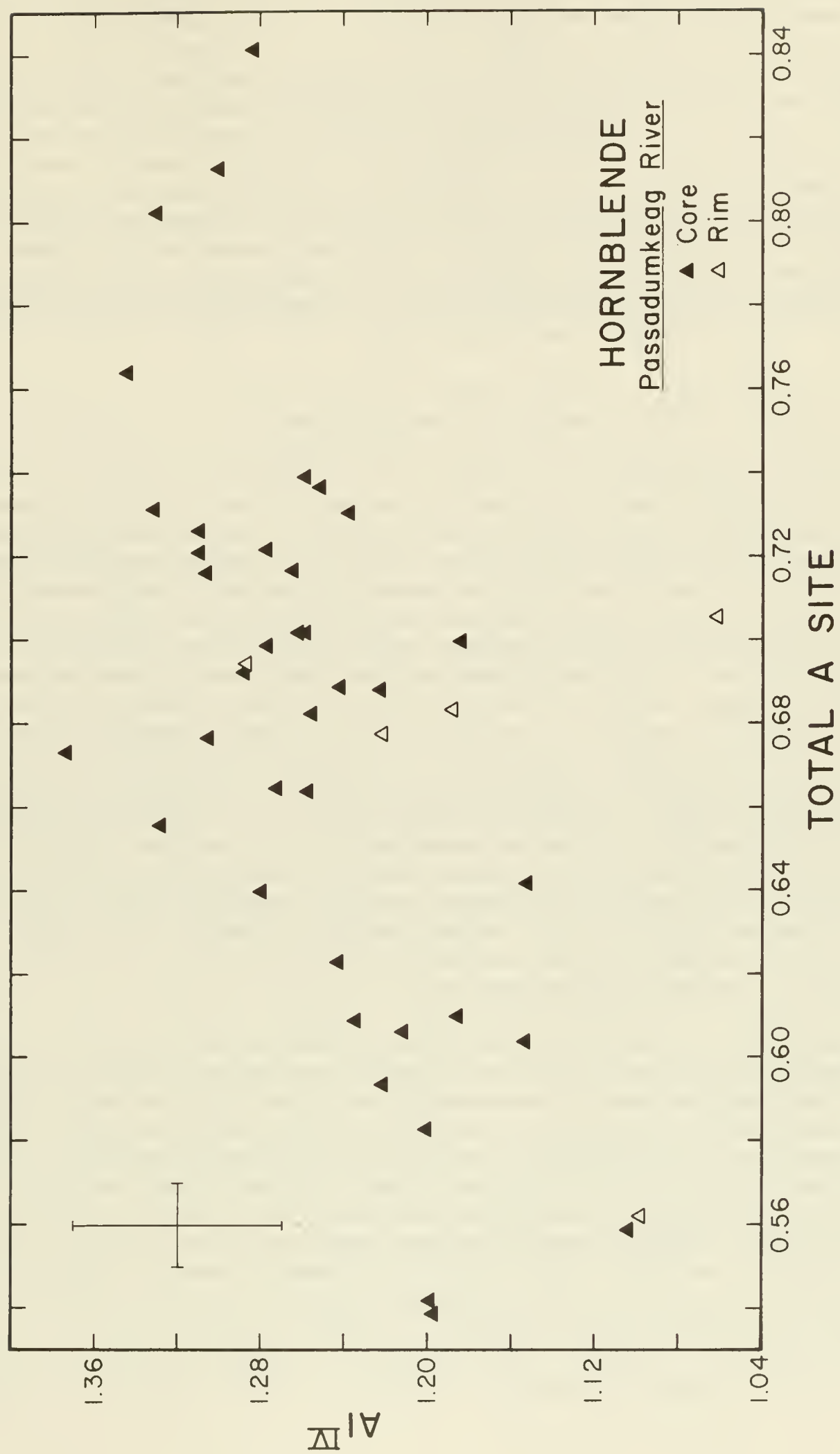


Figure 8 - Representative microprobe analyses of hornblende in the Passadumkeag River pluton, showing Al^{IV} and total A site cation abundances in one-half of the formula unit. Error in determinations shown approximately at 2 σ level.

Cove core facies tends to be highest in anorthite component. In general, no pattern exists toward albite-rich plagioclases. Bulk composition of alkali feldspar is probably about Or_{85-95} based on a few determinations.

Magnetite is often the most abundant opaque phase in both plutons. It is compositionally restricted to titanium-poor varieties (generally less than 1 wt. %) that show no trend from rim to core. Most of it is found as subhedral grains in ferromagnesian minerals, closely associated with ilmenite. Although much of the ilmenite may be secondary, subhedral grains within biotite and in the matrix testify to its primary origin, especially in rocks where it constitutes the only oxide phase. Regardless of pluton, ilmenite is mangiferous and ranges from 3 to 10 wt % in MnO . As in the case of the magnetite, no consistent trend is present between the two plutons and for each facies.

Geologic Interpretation

The Bottle Lake Complex consists of massive, late Paleozoic granitic intrusives emplaced into the greenschist facies rocks of the Merrimack synclinorium. Petrologic similarities within the Complex suggest that the two main granitoid rocks were derived from geochemically equivalent source materials. Both plutons are alike in their essential mineralogy and key index accessory phases (primary sphene, allanite, zircon, apatite). This lack of distinction in the petrography of the Passadumkeag and Whitney Cove plutons is also expressed in their major element variations and abundances by a general overlap in composition and trend. Complete overlap is shown by the lead isotopic composition of feldspar and whole-rocks of both plutons in the Bottle Lake Complex. The range in the isotopic ratios of $^{206}\text{Pb}/^{204}\text{Pb}$ (18.2-18.6), $^{207}\text{Pb}/^{204}\text{Pb}$ (15.6-15.7), and $^{208}\text{Pb}/^{204}\text{Pb}$ (38.2-39.2) is consistent with the interpretation that the source materials for the Whitney Cove and Passadumkeag River plutons are similar (Ayuso and others, 1980b). Preliminary strontium isotopic determinations suggest that both plutons have $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios of about 0.7053, and further support the lead and strontium isotopic similarity of the sources represented by the Bottle Lake Complex.

Source rocks of the Bottle Lake Complex must be dominated by a protolith of igneous character (I-type of Chappell and White, 1974) as suggested by field relations, accessory mineralogy, major and trace element composition, and relatively low initial ratios for lead and strontium. The composite nature of the Complex, the circular or elliptical shape of the plutons, development of an aureole by contact metamorphism, the range in the lithological variety of granitic rocks, and the presence of mafic-rich inclusions in the Complex are in agreement with the field relations outlined by Chappell and White (1974) characteristic of I-type sources. Additionally, the absence of relict aluminous phases (garnet, aluminosilicates, cordierite, or muscovite) and presence of sphene and hornblende are also concordant with such a source. The abundance of magnetite suggests the prevalence of high oxygen fugacities common in I-type materials. Compo-

sitionally, these plutons are also consistent with an I-type source, as they are only slightly peraluminous, and exhibit regular and linear variation in Harker diagrams for major and trace elements for a broad range of silica contents. Finally, typical I-type sources have initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of less than 0.7080, in good agreement with the preliminary value of about 0.7053 for the Bottle Lake Complex.

Lead isotopic studies are inconsistent with an origin directly from mantle sources, granulite rocks in the lower continental crust, or from upper continental crust as represented by sediments exposed near the Bottle Lake Complex (Ayuso and others, 1980b). However, a source consisting mostly of volcaniclastic rocks is supported by the studies of Ayuso and others (1980b) and Loiselle and Ayuso (1980).

The Bottle Lake Complex may represent granitic magmas from a volcaniclastic source, generated by progressively higher degrees of melting and probably emplaced in the same order as their generation. Minimum melting resulted in liquids belonging to the granite of Whitney Cove. Intrusion of the rim facies was controlled by the northeast-trending pre-Late Paleozoic faults in the area; this was followed by intrusion of the core facies which probably represents remobilized material. After intrusion of the Whitney Cove pluton, the fault was reactivated and resulted in the cataclastic zone. The amphibolite rocks in this zone may represent a diabase dike intruded within the fault prior to emplacement of the granite of Passadumkeag River. As in the earlier melting episode, the quartz-monzonitic rocks intruded after the rim facies and they also represent remobilized, refractory material.

Renewed interest in the origin of granitic rocks in eastern Maine is exemplified by studies contrasting field relations (Ayuso and others, 1980a), and geochemical characteristics of the plutons intruding the Merrimack synclinorium (Loiselle and Ayuso, 1980; Ayuso and others, 1980b). Wones (1976, 1980) suggested that the granitic plutons of New England and Sierra Nevada were significantly different, and that they probably represented different tectonic regimes. Recent work by Fyffe and others (1980) on granitoids from New Brunswick further support the absence of typical continent-ocean tectonic regimes. Geochemical study of the granitoids intruding the Merrimack synclinorium in eastern Maine show geochemical gradients unlike those in the Sierra Nevada. The major distinction in granitoids from eastern Maine exists between plutons intruding different tectonic blocks on either side of the Norumbega fault system. Each block could have produced magmas from different processes that are now coincidentally aligned as a result of fault motion. If fault displacement is minimal, production of granitic rocks in the northern Appalachians may indicate a process different from continent-ocean tectonic regimes during the Mesozoic and Cenozoic eras.

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ROAD LOG

Assemble at 8:30 AM in the parking lot of the Lincoln Post Office, just directly across from the First Bible Baptist Church. Most of the trip will follow unimproved, two-lane dirt roads almost exclusively used for hauling timber. Please remember that truckers expect unimpeded access and the right-of-way in these narrow, shoulder-less roads. Because of limited maneuverability and safe-parking in these roads, up to five vehicles will be accommodated on the trip. Participants should bring a bag lunch and expect two short hikes, mostly along trails.

Mileage

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|-----|-----|---|
| 0 | 0 | From Post office, go north on Fleming St. |
| 0.1 | 0.1 | Turn E (right) on Depot St. |

- | | | |
|-----|-----|---|
| 0.2 | 0.1 | Continue past the traffic light. Follow signs to U.S. Route 6. |
| 0.3 | 0.1 | Turn E (left) at the stop sign at intersection with U.S. Route 6, also known as Lee St. |
| 1.0 | 0.7 | Note the abundant, low-lying outcrops of the granite of Center Pond. |
| 1.9 | 0.9 | More exposures of similar granitic rocks. |
| 2.4 | 0.5 | Same type of rocks are exposed here. |
| 2.8 | 0.4 | Continue east on U.S. Route 6 and note on the left side of the road the sign for Schrite's Caribou Pond Cabins. |
| 3.0 | 0.2 | Outcrops on both sides of road belong to the Center Pond pluton. |
| 3.5 | 0.5 | Same type of granitic rocks |
| 4.6 | 1.1 | STOP 1: CONTACT OF COUNTRY ROCKS WITH THE GRANITE OF CENTER POND |

This outcrop shows the prevalent mineralogy and mode of intrusion of the Center Pond pluton. Many of the field relations and petrographic features exemplified by this outcrop are also common in the granite of Passadumkeag River. Both plutons are hornblende-biotite rich, two feldspar granitoids, exhibiting similar accessory mineralogy and textures. Granite-country rock intrusive relationships between the Center Pond pluton and the Vassalboro Fm. are typically transitional in nature and are exemplified in this exposure. The transitional character of the contacts is also common in the Passadumkeag River granite, although knife-edged transitions and others where numerous pegmatites and aplites crisscross the country rock are also well represented. Take a moment to study this outcrop on the opposite side of the road before approaching the outcrop to inspect it more closely. Note that country rock is clearly predominant on one end (eastern) of the outcrop, and that it grades into a transitional or mixed zone on the opposite end (western). On the eastern end of the outcrop, the Vassalboro Fm. exhibits a typical pin-striped, shaly, slightly rusty, and heavily-jointed appearance at the contact with the granite of Center Pond. The purple color developed in these rocks reflects the titaniferous biotite formed during contact metamorphism. Immediately adjacent to the granite-country rock contact, the Vassalboro Fm. consists of pin-striped bands of differing mineralogy. Dark and light bands correspond to biotite and calcite-rich zones, respectively, co-existing with diopside, amphibole, epidote, and quartz.

Large blocks (3 m) of country rock and granite are common within the transition zone and typify the mode of intrusion prevalent in both the Center Pond and Passadumkeag River plutons. Inspection of individual blocks of metasedimentary origin suggest that the pluton intruded in a lit-par-lit manner, disaggregating blocks of various sizes. Aplitic stringers intrude along the bedding planes. Several examples of this relationship are present in this outcrop in elongated blocks about 1 m in size, still exhibiting orientation similar to that in the country rock. This rock shows extreme deformation close to the cataclastic areas and this is expressed by intense recrystallization of quartz, kinking and smearing of biotite, and dislocation of feldspars. Most biotite is red from hematite stain, and it is commonly rimmed by finer-grained opaque minerals. In spite of the intense rotation, grinding, and kinking in plagioclase, these feldspars retain evidence of a large core rimmed by a clear, optically continuous rim. Hornblende still shows subhedral to anhedral outlines with abundant biotite inclusions. Sphene, allanite, apatite, zircon, and opaque oxides are present elsewhere in this pluton; however, they are not conspicuous in this crop. This outcrop also shows cataclastic textures interpreted by Scambos (1980) as part of a NE-trending fault zone through the pluton with an offset of 2 km in a right lateral manner. Note the strongly foliated character of the rock (N80° E, 60-90 SE) as expressed by the feldspars and mafic minerals on the western end of the exposure. This fault zone may also be evident in the granite and country rocks by the abundance of joints oriented generally to the NE, and by a few thin (<5 cm) quartz veins.

Granitic rocks are generally gray in color, lack a porphyritic or quench marginal zone at the contact with the country rock, and preserve a nearly equidimensional character in their fabric. While xenoliths of metasedimentary origin decrease in the granitic rocks away from the transition zone, small clusters (<10 cm) of mafic minerals (biotite + amphibole + sphene) and plagioclase are common throughout the granite. These clusters probably do not represent fully digested country rock as suggested by their mineralogy, texture, and lack of alignment. Because pegmatites and miarolitic cavities are absent and aplitic rocks are uncommon in rocks near the contact, the magma probably intruded without being saturated with volatile constituents. The contact metamorphic effects are constrained to within 1.5 km of the contact. In spite of the cataclastic deformation in these rocks, it is still possible to ascertain that at the time of intrusion, the liquid was probably saturated with respect to all the major mineral phases. This relationship is evident in the nature of the inclusions in alkali feldspars (plagioclase, biotite, amphibole, quartz) and the general subhedral nature of the phases to each other in samples away from the fault zone.

Proceed E on Route 6.

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|------|-----|--|
| 5.6 | 1.0 | Lee township line. Continue E on Route 6. Hornfels in the contact aureole of the Bottle Lake Complex form prominent hills, regardless of lithology and especially in the immediate envelope around the pluton. Marshy land and streams are commonly found on the slopes facing away from the intrusive, while the opposite slope commonly contains exposures of the granite-country rock contact. As we proceed to the east on U.S. Route 6, keep in mind that the row of nearby prominent hills on the right side of the road consists of the contact aureole of the Bottle Lake Complex. |
| 6.2 | 0.6 | Outcrops of Vassalboro Fm. on left side of road. |
| 6.5 | 0.3 | Intersection with South Rd. Continue E on Rt. 6 and notice outcrops on right side of road. |
| 7.1 | 0.6 | More outcrops on left side of road. |
| 7.5 | 0.4 | Jefferson Ski Area Slope is directly ahead. |
| 7.8 | 0.3 | Continue E on Rt. 6 and through the flashing light at the intersection of Rt. 168 and Rt. 6. Lee Academy and Lee Post Office are on opposite sides of street to your left. |
| 8.2 | 0.4 | Outcrops of Vassalboro. |
| 8.6 | 0.4 | Maine Forest Service station to your right. Continue E on Rt. 6. |
| 10.3 | 1.7 | Springfield town line. Continue E on Rt. 6. |
| 10.9 | 0.6 | Note the outcrops across from the AMOCO station. |
| 13.9 | 3.0 | Continue E on Rt. 6. Go by Springfield Post Office on your right, and intersection of Rt. 169 and 170 with Rt. 6. |
| 14.6 | 0.7 | Turn S (right) on Dead River Co. road (Dead River Tree Farm). We are now approaching the contact between the Bottle Lake Complex and country rock, essentially at right angles. Note the progressive increase in size and abundance of granitic rock boulders in the next several miles. |
| 16.2 | 1.6 | Almanac Mountain is directly ahead of us. It is the most prominent topographic feature in the vicinity and may be identified by the firetower and relay antenna on its pinnacle. |
| 16.5 | 0.3 | Continue S and go by Springfield Cemetery on your right. |
| 17 | 0.5 | Outcrops of Pre-Silurian sulfidic schist on the right side of road. |

- 17.1 0.1 Continue S past the intersection of road going to the east. You may be able to see Duck Lake and Bottle Lake directly ahead and on the left side of the road.
- 17.6 0.5 Lakeville town line. Turn left (E) at the intersection with road going to Duck Lake.
- 18.5 0.9 Turn right (S) at fork on road going to Duck Lake.
- 19.2 0.7 Turn left (E) at fork with Getchell Mtn. gravel road. Note the abandoned school at the intersection (Gowell School).
- 19.6 0.4 Cross Getchell Brook
- 20.2 0.6 STOP 2: RIM FACIES OF THE PLUTON OF PASSADUMKEAG RIVER.

Please turn around and try to park as safely and as far off the road as possible. Mr. H. Snow owns this land and needs free access to the road.

This stop exhibits pavements and low ridges forming essentially continuous granitic outcrop to the top of Getchell Mtn. (1041 ft.). Outcrops are generally fresh and very extensive. The road ends within 1500 ft. from the granite-country rock interface and about 4500 ft. from Getchell gravel road. Because of its orientation at right angles to the contact and abundance of fresh outcrop, this traverse constitutes an outstanding opportunity to study the field relations, fabrics, mineralogy and interaction of the granitic and country rocks. Walk up the mountain to the end of the road, and on the way down inspect the outcrops.

Granitic rocks on Getchell Mtn. belong to the rim facies of Passadumkeag River. The incompletely developed rim facies of the pluton may be subdivided into smaller domains on mineralogical and textural grounds. Note in Figure 2 that the rim facies is absent in a large area near the northern contact. All Getchell Mtn. rocks are hornblende-bearing, two-feldspar granites constrained roughly to the area in contact with pre-Silurian (?) sulfidic schist. Schlieren, different types of inclusions, mafic segregations, and irregular felsic masses are common in this traverse. In most cases, the long dimension in the mafic segregations, schlieren, and mafic inclusions strike roughly N 50°E, parallel to the granite-country rock contact. This preferred alignment is also expressed by the long dimension of the alkali feldspars along the traverse. All the granitic rocks show a prominent and consistent NE-trending alignment. Note the presence of seemingly irregular masses of granitic rocks of markedly porphyritic nature surrounded by the more typical rim facies rocks in many of the outcrops. In addition to

these felsic inclusions, aplitic dikes and pegmatitic pods are also numerous and exhibit no difference with proximity to the contact. No consistent orientation is expressed by these dike rocks as they show significant variation in thickness, and contrast with the uniform orientation of the mafic concentrations. The latter attain lengths up to 1 m and show cyclic mineralogic variations from mafic to felsic-rich layers. Mafic inclusions show a great range in size, shape, and mineralogy. They range from the small (5-10 cm), ovoidal, fine-grained, and randomly oriented biotite and plagioclase masses common throughout the pluton, to biotite-rich aggregates, and finally to prominently porphyritic quartz-diorite rocks. Note that in spite of the apparent randomness in orientation of these inclusions, their long axes are commonly oriented roughly to the northeast. Pegmatitic zones up to several meters in thickness are also exposed on the western slopes of the mountain and these consist chiefly of feldspar (15-20 cm), massive quartz (white and rose), pseudo-hexagonal biotite books (5 cm) and sulfide minerals.

Major textural changes are exhibited by granitic rocks near the contact, but unfortunately these are not exposed along the road. A zone of fine-grained and phenocryst-rich rocks is present at the granite-country rock contact; these rocks are associated with numerous aplitic dikes and pegmatites that randomly crisscross the country rock. Study of individual blocks of country rock reveals that the magma also intruded by sending dikes in a lit-par-lit fashion. Gradual changes in color, texture, and mineralogy are evident with distance from the contact. At the contact, the granite is strikingly porphyritic, mafic-poor, and grayish in color. Toward the bottom of the mountain it becomes seriate, richer in mafic minerals and pinker. Amphibole is rare near the contact, but it progressively increases in abundance, grain size and development toward the interior.

Accessory and varietal minerals consist of allanite, sphene, apatite, zircon and opaque oxides and they show a sympathetic increase in abundance toward the interior of the pluton. Intergrowths of biotite with either plagioclase or hornblende are common, especially near the bottom of the mountain. Interfaces where large, inclusion-rich biotite flakes join plagioclase are characterized by a zone of fine-grained albitic plagioclase, quartz, and Fe-Ti oxides. Biotite and euhedral hornblende form intimately related clots suggestive of coprecipitation; however, it is also common to find subhedral biotite inclusions within hornblende. The probable sequence of crystallization near the

granite-country rock contact may be generalized as follows: zircon, opaque oxides (magnetite), apatite, allanite and sphene, followed by biotite, amphibole, plagioclase, quartz and alkali-feldspar. Even most outcrops near the granite-country rock contact show this order of crystallization, except for the absence of hornblende. Myrmekite and graphic textures, together with the formation of chlorite, sericite, hematite, minor epidote, secondary sphene and opaque oxides are also evident through the traverse.

Textural and mineralogic changes are less striking between the rocks along the traverse compared to those near the granite-country rock contact. Granitic rocks of the rim facies progressively attain the mineralogy and texture typical of the quartz-monzonitic core rocks. This type of gradual change toward the core of the pluton is common; however, in the region north of Upper Sysladobsis Lake the transition from felsic rim to core is substantially more abrupt.

Continue W on Getchell Road to Gowell School, retracing route toward Dead River Co. Road.

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|------|-----|---|
| 21.2 | 1.0 | Turn right (N) at the fork. |
| 21.9 | 0.7 | Turn left (W) at the fork. |
| 22.8 | 0.9 | Turn left (S) on Dead River Co. Road. |
| 22.9 | 0.1 | Continue straight, past the road to the firetower on Almanac Mtn. |
| 23.5 | 0.6 | Almanac Mtn. is immediately to our right; note that directly in front are Sysladobsis, Bottle, and Keg Lakes. |
| 23.9 | 0.4 | Contact between the Complex and country rock is approximately here. |
| 24.7 | 0.8 | Continuous outcrops of the core facies of Passadumkeag River pluton on both sides of the road. At this location the rocks are similar to the granitic outcrops near the bottom of Getchell Mtn. |
| 25.1 | 0.4 | Continue straight (S) on this road. Keg Lake is at your left and Bottle and Sysladobsis Lakes are directly ahead. |
| 25.4 | 0.3 | At the fork, take dirt road on the right. The other road is only .3 miles long and ends at Bottle Lake boat ramp. |
| 26.9 | 1.5 | This is a five-way intersection. Take a left on dirt road going south that is a continuation of the Dead River Co. Dump Road. Starting at this point, Sysladobsis Lake will be on the right and we may be able to glance at the lake on the way to the next two stops. This road is the main access for the great number of cottages on the lake. |

- 27.3 0.4 Pug Hole in Sysladobsis Lake is on the right.
- 29.2 1.9 This road transects a hill with abundant exposures of the core facies of the Passadumkeag River pluton.
- 32.8 3.6 Next stop is about 1300' from the fork in the road. Parking and turning around is difficult in this area, especially after a downpour. All vehicles should be turned around and parked on the same road that was used coming into the fork. Occupants should then congregate at the fork for a short hike. Proceed E along the dirt road on the northeast slope of Porcupine Mtn. for approximately 1300' and note outcrops on both sides of road. These are essentially continuous for about 1000' to the east, and extend to the top of Porcupine Mtn.

STOP 3: RIM FACIES OF THE GRANITE OF WHITNEY COVE

This outcrop exemplifies the rim facies of the pluton of Whitney Cove and it consists typically of pink granitic rocks of low color index and seriate texture. Note the abundance of aplites, pegmatites, quartz veins, and granophyre dikes in this area. Also note the relative depletion and limited range of lithologies in these mafic inclusions compared to the previous stop. Felsic dikes are variable in attitude and thickness, exhibiting a tendency to subdivide and criss-cross within a single outcrop. Pegmatitic masses in this area contain coarse-grained feldspars (5 cm) and quartz, and form pods without a consistent orientation.

This exposure is mineralogically similar to most rocks belonging to the rim facies and consists of microcline, plagioclase, biotite, quartz, and accessory minerals. Intergrowths of biotite and plagioclase are reminiscent of the rocks on Getchell Mtn.; however, they are more abundant in this area and the zone of interaction is substantially wider. Myrmekitic and micrographic textures are ubiquitous as is the granulated appearance (mortar texture) enveloping many of the alkali-feldspars. Allanite is conspicuously euhedral, zoned, and very abundant in this area; euhedral opaque oxides are more common than in the rim rocks of the Passadumkeag River pluton and include magnetite, ilmenite, and pyrite.

Return to cars.

Continue N on this road.

- 36.0 3.2 STOP 4: CORE FACIES OF PASSADUMKEAG RIVER PLUTON

Continuous outcrop extends to the top of the hill as pavements, typically with smooth surfaces and covered by a thick lichen carpet. Exposures are also abundant in the opposite direction, close to the shores of nearby Sysladobsis Lake. Note the abundance of mafic inclusions, aplite dikes, and the marked increase in the color index of this rock. In addition to the ovoidal, fine-grained, biotite and plagioclase-rich clusters randomly dispersed in the granitic matrix, note the fine-grained, porphyritic, and hornblende inclusions. The latter are usually the largest mafic inclusions in this facies, and generally show a wide distribution of sizes. A preferred orientation to the northwest is commonly expressed by their long dimensions, although this contrasts with the easterly alignment of the ovoidal clusters and the overall massive fabric in the rock. Mafic-rich bands consisting of abundant biotite and hornblende and striking to the northeast are also abundant in this area; notice that in contrast to the porphyritic inclusions, these bands show an intermingling with granitic material and indefinite boundaries. Mafic content of the porphyritic inclusions varies extensively, sometimes resulting in rocks of lower color index than the surrounding granitic rock. However, note that large euhedral hornblende and alkali feldspar are common to the granitic envelope and porphyritic inclusions. Pods of felsic rocks showing aplitic texture are easily distinguishable from the abundant inclusions of high color index by their mineralogy and amorphous appearance. Few aplitic dikes are exposed in this facies of the pluton; however, several dikes outcrop in this area, varying in thickness, color, and orientation even over small distances. Joints are generally well-developed and in some cases they are strongly stained by hematite.

This outcrop also exemplifies the mineralogical and textural contrast between the core facies of the Passadumkeag River and the Whitney Cove pluton. Amphibole is a common phase in this stop, and it is commonly present as large, euhedral, black prisms. Together with biotite, they impart the characteristic high color index found throughout the core of the Passadumkeag River pluton. Alkali feldspars are prominently displayed, often mantled by plagioclase and enclose all other essential minerals. Two or more growth zones punctuated by oriented mineral inclusions are commonly shown by the larger alkali feldspars. Plagioclase contains abundant biotite and hornblende inclusions; in most cases plagioclase grains are smaller than alkali feldspar and form plagioclase-rich areas that alternate with quartz-rich clusters.

As in the case of biotite, hornblende forms at least two generations distinguishable by texture, size, and abundance. Euhedral hornblende commonly forms large, inclusion-rich (biotite, sphene, apatite, opaque oxides, zircon), randomly distributed grains; however, note that smaller hornblende is also abundant in the fine-grained plagioclase and biotite-rich clusters. As in the Getchell Mtn. case, the relationship between biotite and amphibole is of mutual intergrowths and imply coprecipitation from the magma.

Return to cars.

Continue straight (N) toward the 5-way intersection.

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| 40.2 | 4.2 | Back at the 5-way intersection. Turn right on the same dirt road used coming in and proceed toward paved road. |
| 41.7 | 1.5 | Turn N (left) at the paved Dead River Co. road. |
| 44.4 | 2.7 | Turn W (left) on gravel road leading to firetower on Almanac Mtn. |
| 46.0 | 1.6 | STOP 5: CONTACT BETWEEN PASSADUMKEAG RIVER PLUTON AND COUNTRY ROCK |

Park in the small parking lot, find a comfortable place to relax and eat lunch.

Extensive outcrops of the contact relations between country rock and the Passadumkeag River pluton are exposed on the southern slopes of Almanac Mtn.; these exposures are essentially continuous to the marked break of slope that envelopes the mountain. We will inspect an area about 100 m. (S20°W) from the firetower. Please follow the trip leader as we proceed toward these outcrops and stay with the group. This stop serves as an example of the type of relationships associated with the granite-country rock contact. As in the case of the pluton of Center Pond, the intrusive contact is expressed along a transition zone characterized by intermingling of country-rock and granitic blocks and lit-par-lit disaggregation; however, small (<10 cm) pegmatites, pods and aplitic stringers of variable thickness are also abundant, and these randomly crisscross the country rock resulting in a brecciated appearance.

Several types of granitic rocks are evident near the granite-country rock contact in the vicinity of Almanac Mtn. These range from strongly foliated, equidimensional, medium-grained, mafic-poor rocks, to poorly foliated, porphyritic, and more mafic. A common attribute of all granitic rocks is, however, that foliation parallels the

intrusive contact. Mafic inclusions are relatively scarce, as compared to the core facies, and they apparently consist mostly of small (< 10 cm) ovoidal, plagioclase and biotite clots.

Granitic rocks immediately adjacent to the contact are notably poorer in mafic minerals compared to the core of the pluton. Feldspars, biotite and quartz are often subhedral even at the contact, and suggest that the granitic magma was already saturated with respect to them at this intrusive level. Note, however, that amphibole is rarely present in these rocks, and that the small concentrations of black minerals are dominated by commonly euhedral biotite.

Hornfelsic rocks are generally rusty on weathered surfaces and purple in fresh cut. Many of the outcrops on Almanac Mtn. show evidence for remobilization and plastic deformation probably related to intrusion of the Bottle Lake Complex plutons.

- | | | |
|------|-----|---|
| 47.6 | 1.6 | Turn N (left) on Dead River road going toward intersection with Rt. 6. |
| 50.8 | 3.2 | Turn E (right) on Rt. 6 going toward the town of Topsfield. |
| 51.6 | 0.8 | Carroll Townline, continue E on Rt. 6. |
| 55.2 | 3.6 | Bowers Mtn. is at your right and consists of both hornfels and granitic rocks of the Passadumkeag River pluton. Hornfels are well exposed on the top and continue down to the southeast limb of the mountain. |
| 57.2 | 2.0 | Pre-Silurian (?) Sulfidic schist outcrops. |
| 57.9 | 0.7 | Kosuth townline, continue E on Rt. 6. |
| 62.2 | 4.3 | Continue E on Rt. 6, past the dirt road going to Maine Wilderness Canoe Basin campground on Pleasant Lake. |
| 64.2 | 2.0 | The lake directly ahead and on the left side of the road is East Musquash Lake; East Musquash Mtn. is the highest mountain in the vicinity and it is identified by the firetower on the top. |
| 64.8 | 0.6 | Turn S (right) at the intersection with dirt road, sometimes known as Amazon Road. Note the two signs at the entrance identifying it as property of Georgia-Pacific Corp., and also giving directions to the cottages on the shores of Pleasant Lake. |

- 68.7 3.9 Continue S on Amazon Rd. Orié Lake is at your left and also notice that East Musquash Mtn. is in the background.
- 69.4 0.7 Pavements of country rock are almost continuously exposed near the road, grading into granitic rocks of the rim facies of the Whitney Cove pluton. Granite-country rock contact is constrained within 200 m. Granitic rocks are foliated parallel to the contact.
- 69.9 0.5 Continue S on Amazon Rd.; trail to the left goes to West Musquash Lake.
- 70.4 0.5 Pavement of rim facies of granite of Whitney Cove.
- 72.1 1.7 Continue straight on Amazon Rd.; trail to the right ends at Pleasant Lake.
- 73.2 1.1 STOP 6: CATACLASTICALLY DEFORMED RIM FACIES ROCKS OF THE GRANITE OF WHITNEY COVE

This outcrop shows the cataclastically deformed granitic rocks in the fault zone transecting the pluton of Whitney Cove and continuously exposed in a belt approximately 1.5 km wide from Orié Lake to Junior Lake. The fault zone is cut by the pluton of Passadumkeag River on its southwest extension, and it may connect with the north-trending fault through the country rocks mapped by Ludman (1978a). The Norumbega fault system and the fault zones through the Bottle Lake Complex and Center Pond plutons show the same northeast trending orientation, and probably represent rejuvenated, major tectonic lineaments in the region.

Note the markedly splintery, jointed, greenish appearance of this outcrop as a result of fault motion after intrusion of the Whitney Cove pluton. Cataclastic outcrops are commonly cut by numerous epidote and quartz-filled dikes, striking generally in the northeast direction, but commonly exhibiting substantial variation in attitude. The fault zone in this area is characterized by granitic rocks that show brittle deformation confined to narrow zones. Stresses are apparently dissipated in the immediate vicinity of the zones of maximum deformation, and are exemplified by spindle-shaped quartz grains within the epidote-rich dikes. Right lateral motion along many of the east trending joint sets is generally confined to less than 50 cm of displacement, and it is best displayed in the transected aplitic dikes. Even the mylonitic dikes are displaced by a few centimeters by this set of joints. Although the fault belt is delineated by a zone of abundant cataclastic rocks in this area, this zone is also expressed by ubiquitous epidote-filled dikes throughout the Whitney Cove pluton; this is especially applicable to the area between this fault zone and the Norumbega fault system.

Note the pervasive and massive alteration recorded by these rocks. Even those feldspars away from the cataclastic dikes are crisscrossed by thin, epidote-rich veins, and show marked alteration effects; biotite is completely chloritized and forms random clusters in the matrix. Granitic textures are obliterated near the epidote-filled veins and are characterized by feldspars showing gradual disintegration, shearing, rotation, and offset; note that the feldspars are immersed within plastically deformed bands of recrystallized quartz and biotite. Many of the exposures in this area are characterized by aplitic dikes that are commonly lined by silicified dikes dotted by fine-grained sulfide minerals, up to 10 cm in thickness, and oriented to the northeast. Foliation in granitic rocks is often well-developed and in good agreement with the orientation of epidote-filled veins.

- | | | |
|------|-----|---|
| 74.3 | 1.1 | Cross Rainey Brook |
| 75.0 | 0.7 | Continue straight on Amazon Road; trail to the right goes to a state campground along the south shore of Scraggly Lake on Hasty Cove. |
| 75.5 | 0.5 | Note the abundance of cataclastically deformed granitic boulders in this area. |
| 75.9 | 0.4 | Pavement of rim facies of Whitney Cove pluton rocks. |
| 76.4 | 0.5 | Continue straight at the intersection. |
| 76.7 | 0.3 | Continue straight at intersection. |
| 77.3 | 0.6 | Continue straight at intersection. |
| 78.1 | 0.8 | Pug Lake in Junior Bay is at your right. |
| 79.3 | 1.2 | Continue straight at the intersection; road to the right goes to Whitney Cove in Grand Lake. |
| 79.5 | 0.2 | Exposures of the porphyritic core facies of the Whitney Cove pluton. |
| 80.0 | 0.5 | STOP 7: PORPHYRITIC FACIES OF THE GRANITE OF WHITNEY COVE PLUTON |

Outcrops of this facies are best exposed along the shores of Whitney Cove in Grand Lake. Contacts between the core and rim facies of this pluton are rarely sharp, but rather exhibit a gradual and progressive transition into the rim facies where observed in the large outcrops near Pork Barrel Lake.

Distinction between the rim and core facies of the pluton depends on the development of strongly porphyritic and fine-grained character of the core rocks. Mafic minerals are predominantly biotite, and as in the rim facies, hornblende is absent; also in common with the rim facies, biotite locally forms conspicuously euhedral, and larger pseudo-hexagonal plates. Ovoidal clusters of fine-grained biotite and plagioclase are generally small (< 5 cm) but more abundant than in the rim facies. Metasedimentary and quartz-dioritic inclusions are relatively rare.

Phenocryst mineralogy is dominated by the feldspars, but quartz and biotite are also represented as subhedral grains; alkali feldspar is clearly predominant in size, abundance and development over inclusion-rich (biotite) plagioclase. Alkali feldspar encloses all other phases and it is sometimes decorated by them along growth zones. In samples where the porphyritic texture is best developed, plagioclase is nearly absent from the rock, and euhedral alkali feldspar is abundant; in spite of their idiomorphic nature, most alkali feldspars show minor development of scalloped edges resulting from embayment by the fine-grained matrix. Rapakivi texture is typically present but it is never abundant.

Biotite is the first major phase to crystallize as evidenced by the abundant pseudo-hexagonal grains and common inclusions within plagioclase; biotite rims of these larger grains are often ragged and intermingled with feldspars. A second generation of biotite is generally ragged in appearance, fine-grained, and forms an interlocking net around all other phases. Accessory and minor phases are represented by allanite, apatite, sphene, opaque (magnetite and ilmenite) and sulfide minerals. Crystallization of biotite, plagioclase, quartz and alkali feldspar followed the accessory minerals, and this sequence probably represents the generalized order of appearance. The only common occurrence of muscovite in the core facies is in wide (>50 cm) aplitic dikes that are numerous in the area surrounding Whitney Cove.

Return to cars.

Continue on Amazon Road.

- | | | |
|------|-----|----------------------------------|
| 82.1 | 2.1 | Outcrops of porphyritic granite. |
| 83.0 | 0.9 | Note Georgia Pacific sign. |
| 83.1 | 0.1 | Cross stream. |

- 84.2 1.1 Granitic rocks of core facies.
- 84.7 0.5 Nice pavements of granitic rocks of Whitney Cove granite.
- 85.2 0.5 Extensive outcrops on the right, almost continuous.
- 85.7 0.5 STOP 8: RIM FACIES OF THE GRANITE OF WHITNEY COVE

Large exposures of this facies are common on most hilltops, and these are exemplified by the massive outcrops exposed on the Pineo Mountains, directly to the south and by the outstanding exposures common on the cliffs and islands of Grand Lake. In spite of the relatively abundant outcrop coverage, most rocks are deeply weathered and difficult to sample appropriately. This stop is close to the internal contact of the rim and core facies, as well as to the unexposed eastern granite-country rock contact. Note however, that this rock retains an overall massive appearance except where the long dimensions of feldspar suggest a preferred orientation to the north. Metasedimentary inclusions are progressively more numerous in this rock with proximity to the contact. However, in addition to these inclusions, note those of quartz-dioritic composition and restitic texture exhibit markedly inhomogeneous distribution even at the scale of the outcrop.

Although most rim rocks show remarkable textural homogeneity, a few pods of variable size are reminiscent of the finer-grained core facies. Epitote-filled fractures and joints are common in many of these exposures and serve as the principal fractures from which numerous secondary fractures originate. Close to these fractured areas, quartz has strongly undulatory extinction and both biotite and plagioclase show kink bands. Note the thick (~60 cm) gray, aplitic dike showing indistinct margins with the granitoid and remarkable zonation from rim to core. Mafic minerals, chiefly biotite, are concentrated at the margin with the granite and are replaced by a felsic band followed by a pegmatitic seam (felsic and mafic minerals) toward the central portion.

Contrast the color, texture, and mineralogy of this outcrop with the previous stops. These rocks are undisputably pink, equidimensional in texture and consist of subhedral biotite (apatite + zircon + magnetite + ilmenite) in discrete clusters in the matrix, usually in close relationship with plagioclase. Feldspar mineralogy is still overwhelmingly dominated by alkali feldspar; to, mostly as subhedral and typically deeply embayed grains. Plagioclase, biotite and quartz are commonly found as inclusions in alkali feldspar, typically in random

orientations without crystallographic control. Rapakivi textures are not abundant, but where present, they consist of peculiarly skeletal and embayed alkali feldspars, enveloped by a wide rim of plagioclase; inclusions of biotite and plagioclase are common in these alkali-feldspar cores. Idiomorphic, gray quartz clearly embays and predates not only alkali-feldspar but also some plagioclase; clusters of quartz mosaics tend to form an essentially monomineralic and interconnected net that intermingles with feldspar clots. Plagioclase shows polysynthetic twinning and a generally euhedral, homogeneous core enveloped by a progressively more albitic rim. Together with biotite and Fe-Ti opaques, plagioclase exhibits peculiar intergrowths and deep embayments indicative of a reaction relationship.

- | | | |
|------|-----|--|
| 88.0 | 2.3 | Cross stream and continue straight. |
| 88.5 | 0.5 | Continue straight at intersection. |
| 88.9 | 0.4 | Pavements of granite of Wabassus Lake. |

STOP 9: (OPTIONAL) WABASSUS LAKE GRANITE

This medium-grained granite is found only within the Norumbega fault zone. It is finer-grained than either the rocks of the Whitney Cove pluton to the north or the Cranberry Lakes pluton to the south. It is more altered than either of those plutons, and has fewer mafic minerals. It does not correspond to either of the three plutons south of the Norumbega fault zone (Lucerne, Lead Mountain, Cranberry Lakes) or to the Whitney Cove pluton north of the fault. As a result, we have been unable to obtain a meaningful estimate of movement on the Norumbega fault zone.

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|------|-----|---|
| 92.2 | 3.3 | Turn E (left) at intersection with paved road and continue to the intersection of U.S. Rt. 1. |
| 93.4 | 1.2 | Cross Big Musquash and continue E. |
| 95.7 | 2.3 | Continue E (straight). Road at your right goes to Peter Dana Pt. |
| 97.3 | 1.6 | Turn N (take left) at intersection with Rt. 1 and continue N to Presque Isle. |

END OF TRIP

GEOLOGY AND PETROLOGY OF IGNEOUS BODIES WITHIN THE KATAHDIN PLUTON

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Introduction

The Katahdin Pluton is a prominently placed plutonic body within a discontinuous bimodal belt of plutonic rocks extending northeastward from Rangeley, NW Maine, toward south of Presque Isle, NE Maine (Fig.1, Fig.2, see also Fig.1 of Ayuso, Wones -this guidebook). This belt, named here as the Greenville Plutonic Belt (after the town of Greenville approximately in the middle of the belt), consists of a series of NE-SW elongated plutons ranging in size from a few km to 70 km and ranging in composition from ultramafics (dunites of Moxie Pluton: Visher, 1960) to granites (IUGS classification) with under-representation of intermediate compositions.

The Greenville Plutonic Belt is defined here as a separate geological entity principally due to its unique tectono-petrogenetic characteristics. First, the belt marks an abrupt NW termination of large-scale Acadian plutonism within the Appalachian Orogen. Second, it parallels a belt of penecontemporaneous acidic volcanic rocks preserved in several synclinal structures immediately to the northwest (Piscataquis Volcanic Belt, Rankin, 1968 -Fig.2). Third, within the belt there are several large gabbroic bodies (Fig.2) which point to a mantle-crust coupling of igneous activity during the Acadian orogenic event. Finally, the belt cuts across several metamorphic isograds, from a subchlorite zone in the NE of Maine to the sillimanite zone in the SW (Fig.3). Since the exposures of sillimanite grade are a reflection of later, postorogenic isostatic rebound due to continuing erosion, it is possible, by assigning reasonable lithostatic pressures to each of the isograds to make an estimate of the magnitude of the rebound. These calculations indicate that the present surface of the Greenville Plutonic Belt resulted by a differential regional "uplift" (tilt) of 5% grading linearly upward from NE to SW. Alternatively, the surface exposures in the Rangeley region represent conditions and environments which, immediately following the Acadian orogenesis, existed at a depth 8 km greater than the conditions of similar exposures at the Katahdin Pluton. This difference gradually diminishes to zero toward the Katahdin Pluton. The Katahdin Pluton is located further to the NE where it is situated between the shallower quartz diorite of the Moxie Pluton and the volcanics of the Traveler Rhyolite. The volcanics of Traveler Rhyolite are included in the upcoming general discussion of the Katahdin Pluton. This is for reason that the Traveler Rhyolite, as it will be shown later, is comagmatic with the Katahdin Granite.

The area of interest to this report was mapped in the early 60's by D. Rankin and A. Griscom (Ph.D. theses - Harvard University) and in the early 70's by the author (Ph.D. thesis - M.I.T.). D. Rankin mapped the area of Traveler Rhyolite; A. Griscom the area of Katahdin Pluton; and R. Hon studied both units petrologically, mineralogically, and geochemically.

Figure 1. Index Map

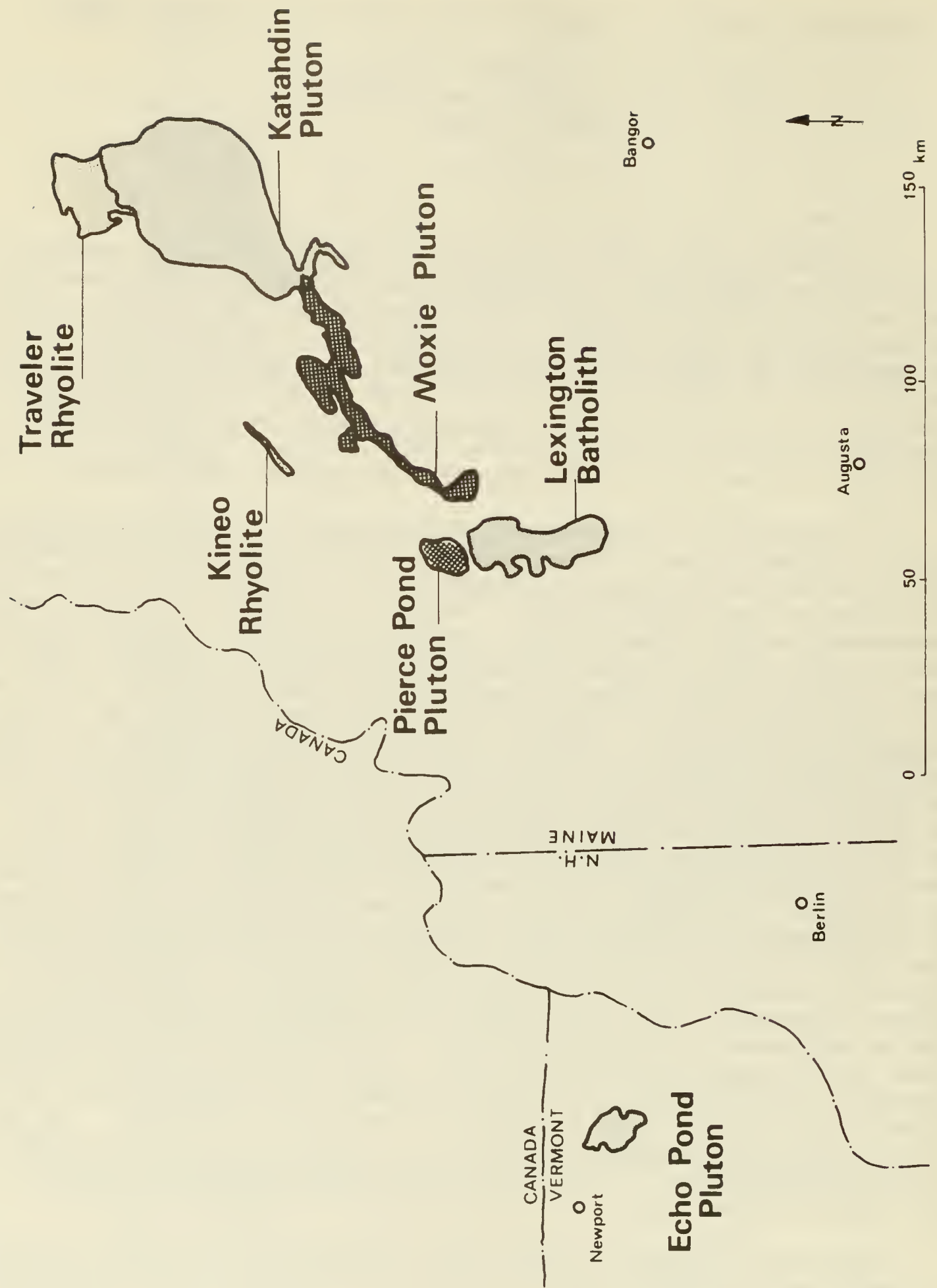
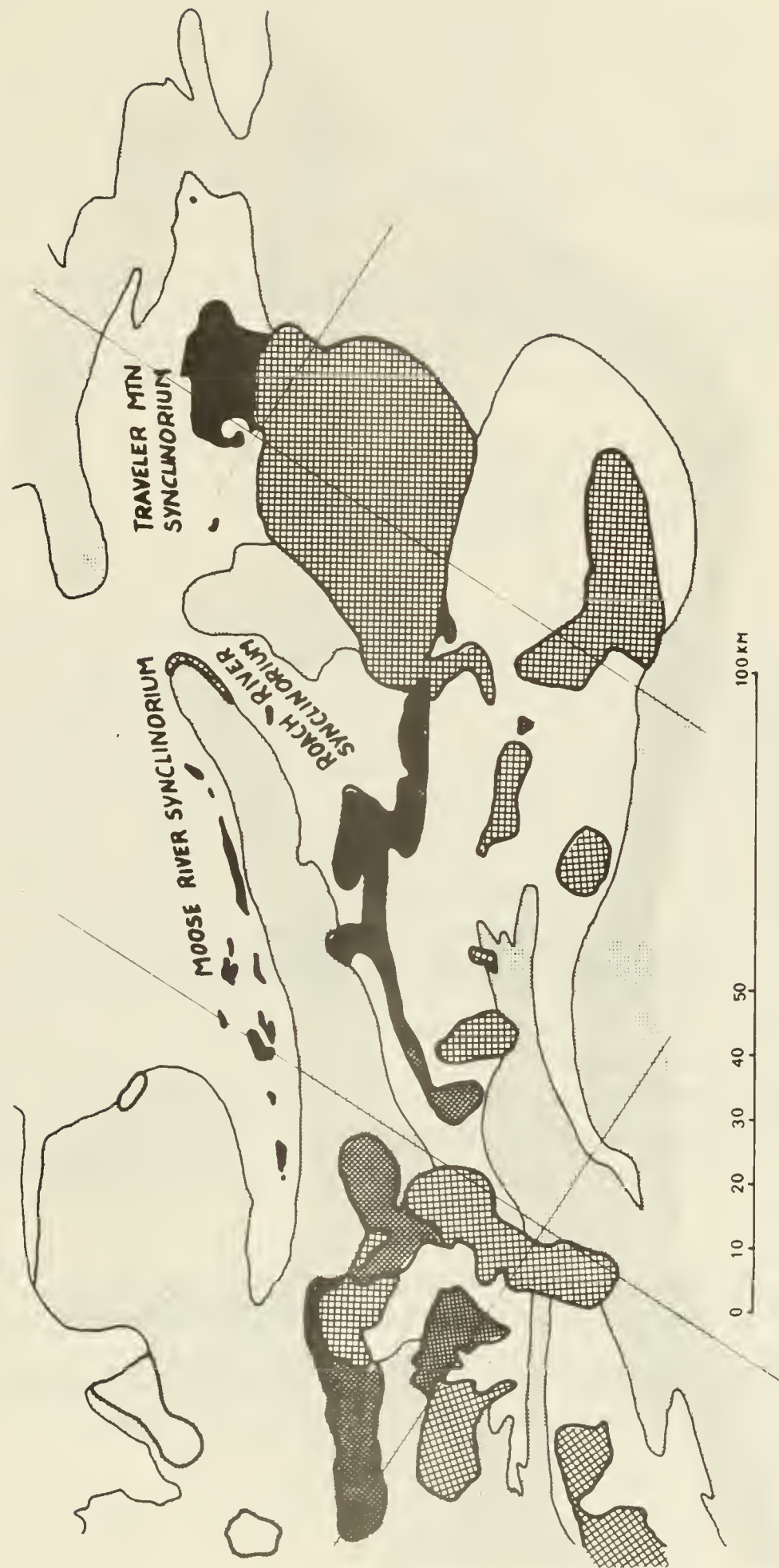


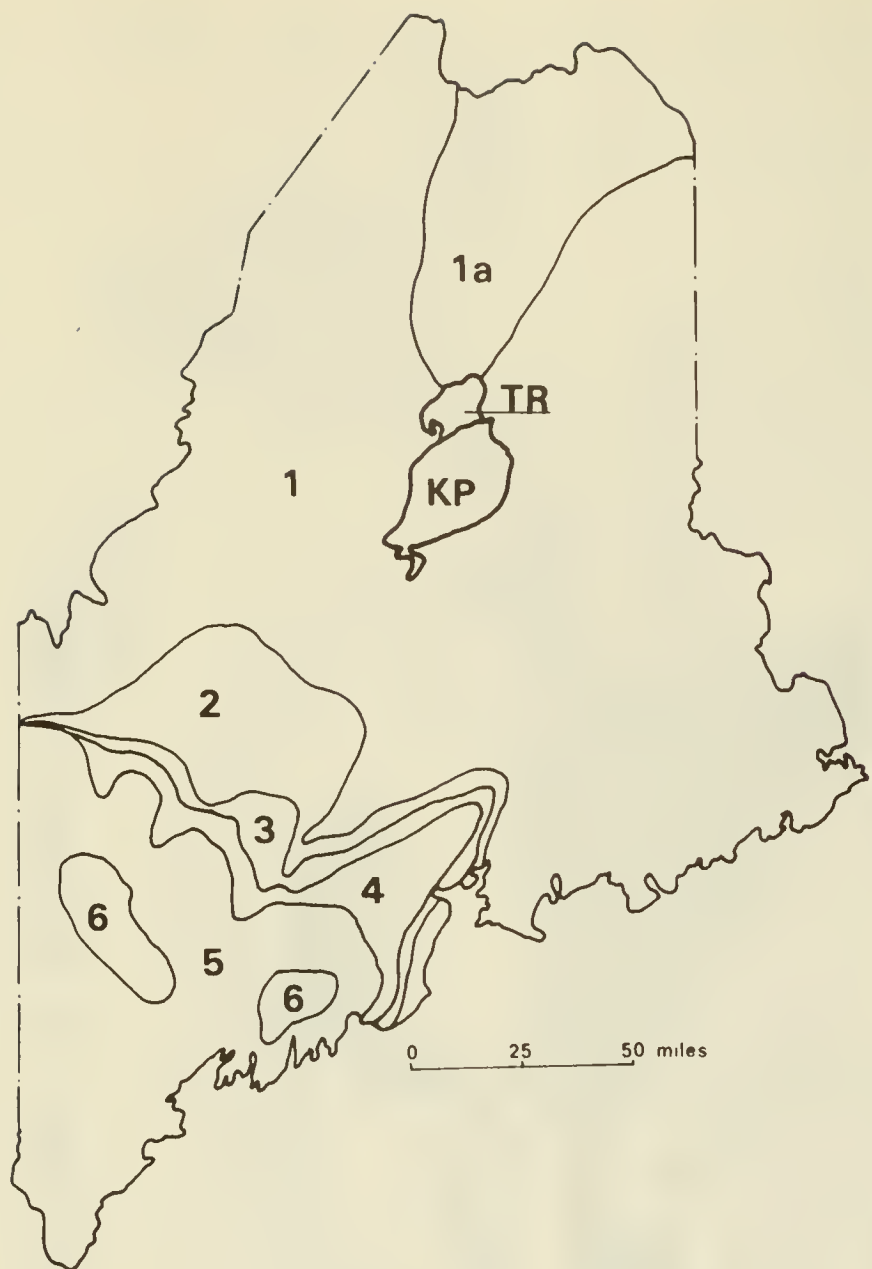
Figure 2. Greenville Plutonic Belt in western Maine.



- Legend:
- 1- Piscataquis Volcanic Belt;
 - 2- Mafic plutons of Greenville Plutonic Belt;
 - 3- Granites of Greenville Plutonic Belt;
 - 4- Other L. Devonian granites;
 - 5- Pre-Devonian sediments;
 - 6- Devonian sediments.



GENERALIZED GEOLOGIC MAP OF WESTERN MAINE



**INDEX MAP AND
METAMORPHIC ZONES
(GEOLOGIC MAP OF MAINE, 1967)**

Figure 3. Metamorphic zones in Maine.

Legend: 1a- subchlorite zone; 1- chlorite zone;
2- biotite zone;
3- garnet zone;
4- staurolite zone;
5- sillimanite zone;
6- sillimanite & orthoclase zone.

NOTE: The metamorphic zones in S. Maine are omitted for clarity.

TRAVELER RHYOLITE - KATAHDIN PLUTON IGNEOUS COMPLEX

The Traveler Rhyolite - Katahdin Pluton igneous complex consists of a 3.2 km thick sequence of rhyolitic volcanics of the Traveler Rhyolite (TR) Rankin, 1968; (Rankin, this guidebook) and a shallow, composite, predominantly granitic Katahdin Pluton (KP). Much of the area of KP (over 95%) is underlain by Katahdin Granite (KG) with the remaining area divided between a stock of Horserace Quartz Diorite (HQD) and two smaller intrusions of Debsconeag Granodiorite (DGD). All three of these smaller bodies (HQD, DGD) intrude Katahdin Granite along the NW-SE trending West Branch Penobscot River fault system (Fig. 4) suggesting that the intrusions were tectonically controlled.

Petrological and geochemical evidence suggests that within the TR-KP complex there existed two genetically independent magmatic episodes. The slightly younger episode includes voluminous ash-flow deposits of TR and subsequent crystallization of KG, whereas the slightly younger episode includes the HQD and DGD.

TRAVELER RHYOLITE - KATAHDIN GRANITE SERIES

Traveler Rhyolite

The Traveler Rhyolite (TR) is an almost rectangular (20 km x 12 km), largely tectonically sunken block of topographically prominent Lower Devonian volcanic rocks, occurring near the NE extension of the Katahdin Pluton. The dominant rock type of TR is a welded ash-flow tuff with varying degrees of compaction (up to 1:20) and variable phenocrystic content. Based on the latter, in particular the presence or the absence of quartz, the volcanic pile is divided by Rankin, 1968, into the basal Pogy Member (quartz present) and the overlying Black Cat Member (quartz absent). Other phenocrysts include plagioclase, ferroaugite, biotite, fayalitic olivine, and opaques. By applying various geothermometers and geobarometers, the phenocrystic assemblage indicate that the phenocrysts of the Black Cat Member form an equilibrium assemblage at $T=800^{\circ}\text{C}$, $P(\text{H}_2\text{O})=1100$ bars and $f(\text{O}_2)$ near the fayalite-magnetite-quartz buffer curve. The temperature for the Pogy Member is approximately 40° lower. The character of the contact between TR and Katahdin Granite is clearly intrusive, documented by numerous apophysal injections of Katahdin Granite into the volcanics and by a narrow zone (100 m) of recrystallized metarhyolites.

Katahdin Granite

Katahdin Granite underlies over 95% of the 1350 km^2 large Katahdin Pluton. The pluton is of a subcircular shape, conspicuously elongated in the NE-SW direction parallel with the direction of the above estimated 5% postintrusive regional tilt. Thus the SW extension of the pluton represents a section which was originally approximately 3.5 km deeper than the corresponding section at

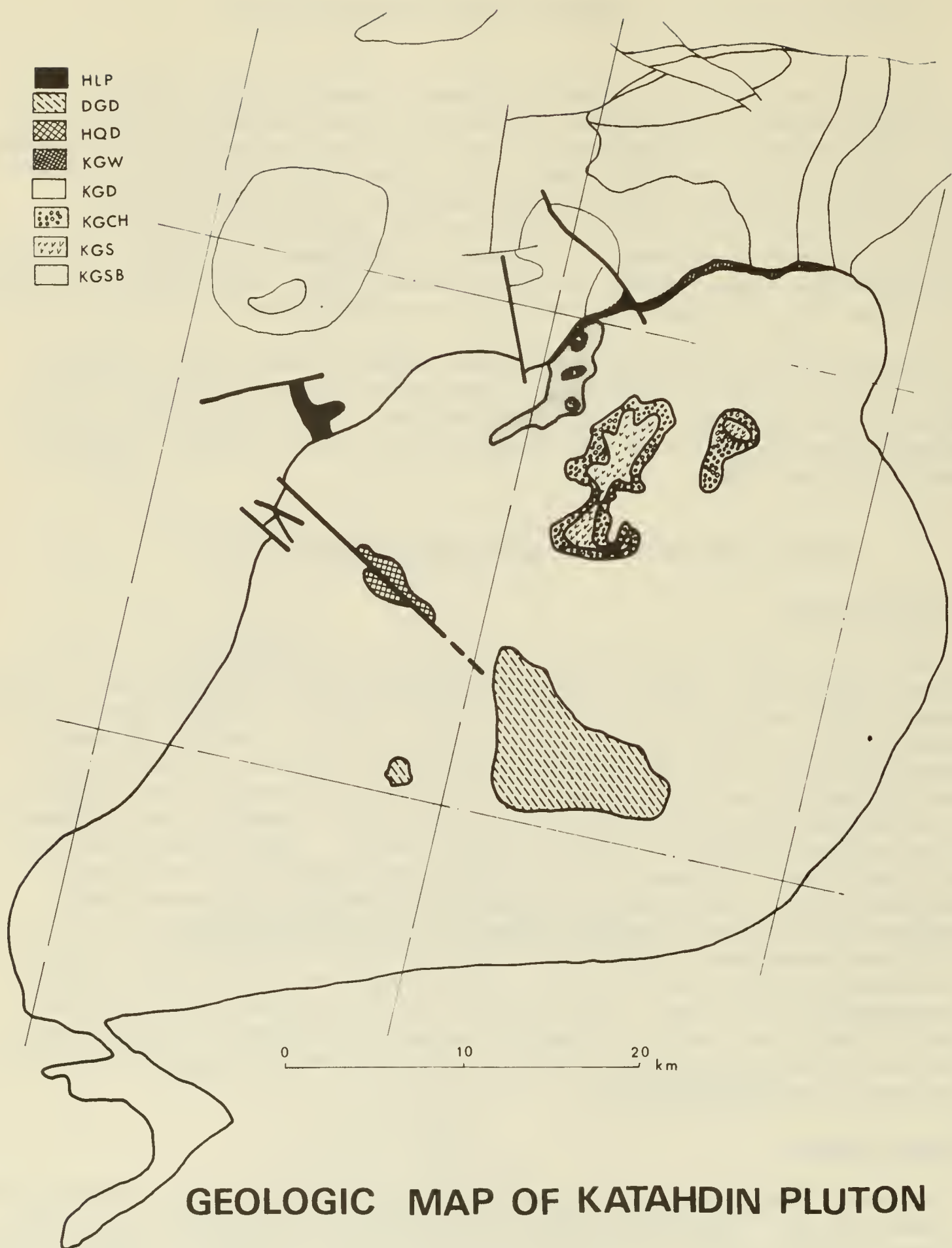


Figure 4. Geologic Map of Katahdin Pluton.

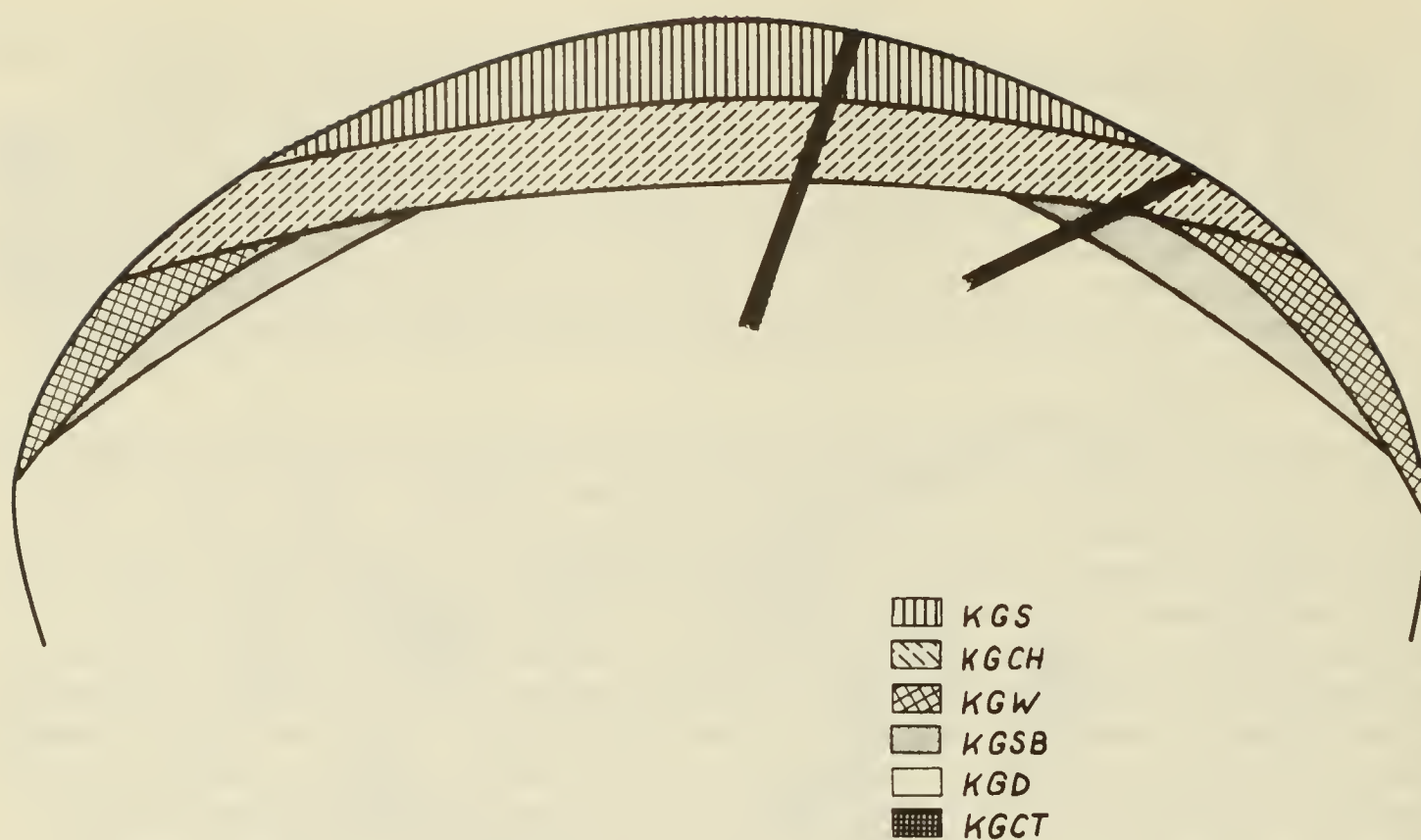
HLP - Harrington Lake Porphyrg
 DGD - Debsconeag Granodiorite
 HQD - Horserace Quartz Diorite
 KG - Katahdin Granite

the NE. This suggests that the three-dimensional form of the Katahdin Pluton is a flat laccolith not yet fully unroofed along the contacts with the Traveler Rhyolite. This is further supported by the observed contact dips and the lack of gravity anomaly associated with the pluton (Kane, pers. comm.). Size estimates of the Katahdin Pluton laccolith are approximately 40 km in diameter and 5 km thick. The general shape of the laccolith is also suggested by the spatial distribution of various textural facies of Katahdin Granite. Fig. 5 illustrates the idealized distribution of these textural varieties within the pluton.

The core of the pluton is formed by a massive, structureless medium-grained biotite granite - the Doubletop facies (Fig.5). This facies grades upward into the Chimney facies through a development of granophyric intergrowths which becomes progressively more and more pronounced and finer with higher elevations, ultimately accounting for more than 60% of the rock. At higher elevations, toward the summit of Mt. Katahdin, the granite becomes also miarolitic, again progressively more prominent toward higher elevations and concurrently changing character from filled vugs into open vugs (Summit facies). Textural variations across the horizontal profile, toward the side contacts are characterized by an evolution of subporphyritic and rapakivi texture, the South Brother facies. This facies, near the side contacts becomes aplite with occasional "nests" containing abundant linings of black to dark blue tourmaline (Wassataquoik facies). The observed thickness of the "outer zone" facies varies from zero to a minimum of 450 m, which is best explained by local differences in the cooling rates.

Aplitic dikes occur throughout the pluton and their frequency seems to be independent of elevation. The frequency of pegmatitic dikes is however inversely proportional to the frequency of miarolitic cavities. The pegmatites are absent at higher elevations but become more frequent toward the core of the pluton.

Also of interest is the relative erosional resistance of the "outer zone" granites. The Summit facies, by its extreme resistance against weathering provides an effective anti-erosional capping at higher elevations, topographically identified in the northern part of the pluton as an elevated plateau contoured by sharp edges, deep valleys and glacial cirques. The plateau, surrounding near its southern edge the highest point in Maine (Mt. Katahdin: Baxter Peak - 5267 ft.), dips gently toward the NNE from 4600 ft. to 2800 ft. over a distance of 8.8 km. The calculated mean slope of the plateau (6%) correlates closely with the estimated postorogenic tilt in this region (5% - see Introduction). It is believed that the inclination of the plateau is not intrinsic to the pluton but rather a result of later isostatic crustal re-equilibration. Wherever the anti-erosional "shield" of the outer zone granites erodes away, the granites of the Doubletop facies are subject to a relatively quick erosion due to their poorer resistance. This explains the existence of steep hillsides around the exposures of the Summit facies and the topographic lows in the remaining part of the pluton.



DISTRIBUTION	FACIES OF KAT. GRANITE		TEXTURE
<u>Core</u>	Doubletop	KGD	Granitic
<u>Outer zone</u>	Chimney	KGC	Granophyric
	Summit	KGS	Granophyric and Miarolitic
	South Brother	KGSB	Subporphyritic with Rapakivi Overgrowth
	Wassataquoik	KGW	Aplogranitic
<u>Porphyritic dikes</u>	Cathedral	KGCT	Porphyritic with Aplitic Groundmass

Figure 5. Schematic distribution of textural facies of Katahdin granite within the Katahdin Pluton Laccolith.

Irrespective of its textural variations, the Katahdin granite is a homogeneous, massive, medium to fine-grained biotite granite of constant chemistry and mineralogy (Fig.6): alkali feldspar OR70:AB28.5: AN1.5 (34%), zoned plagioclase AN34 to AN25 (26%), quartz (34%), biotite (6%), and accessory apatite, allanite, zircon, tourmaline, opaques. There is no known occurrence of muscovite, garnet, or sphene. Various geothermometers and geobarometers indicate temperature of crystallization at $T=710^{\circ}\text{C}$, $P(T)=P(\text{H}_2\text{O})=1200$ bars, and $f(\text{O}_2)$ near the fayalite-magnetite-quartz buffer curve. The pressure estimate compares well with the estimated 1100 bars for the phenocrystic assemblage of the Traveler Rhyolite, suggesting that the KG represents a solidified magma chamber underneath the Traveler volcano.

HORSERACE QUARTZ DIORITE - DEBSCONEAG GRANODIORITE SERIES

Horserace Quartz Diorite

The Horserace Quartz Diorite (HQD) forms a small elongated stock intruding the Doubletop facies of Katahdin granite in the western part of the pluton (Fig.4). The stock is about 5.5 km long, and at its maximum 2.0 km wide with outcrops on both sides of the West Branch Penobscot River. The HQD intrusion parallels a prominent fault system extending NW-SE for a minimum of 30 km from Chesuncook Lake to the center of the KP. On both sides of the HQD stock, also parallel with the fault system, are numerous lamprophyre-like dikes which are mineralogically and chemically identical with the HQD. The ascent of the quartz dioritic magma was therefore tectonically controlled and the fault remained active even after the intrusion solidified.

From aeromagnetic data, Allingham, 1960, deduced that the intrusion is symmetrical with contacts dipping outward at about 55° to a depth of at least 3 km. Projecting the contacts upward, the upper contact can be estimated to have been about 1.5 km above the present exposures. The contacts with the surrounding KG are consistent with the model of a forceful tectonically controlled intrusion. Brecciation of the granite and granitic inclusions within the HQD are commonly observed near the contacts.

The HQD intrusion is concentrically zoned with a composition of biotite-amphibole quartz diorite at the margins grading inward into amphibole-bearing biotite granodiorite. Color index varies accordingly from about 25 to about 7 in the center. The principal mineral components (Fig.6) are plagioclase (andesine to oligoclase), quartz, alkali feldspar, hornblende, biotite, and accessory apatite, sphene, opaques and fine primary "droplets" of sulphides (probably pyrrhotite and chalcopyrite almost exclusively in the cores of amphiboles). Fig.6 shows compositional variations plotted on the IUGS classification triangle as well as variation diagrams suggesting that the principal cause of differentiation involves simultaneous removal of plagioclase and amphibole. Estimates of $f(\text{O}_2)$ yield values above or near the Ni-NiO buffer curve.

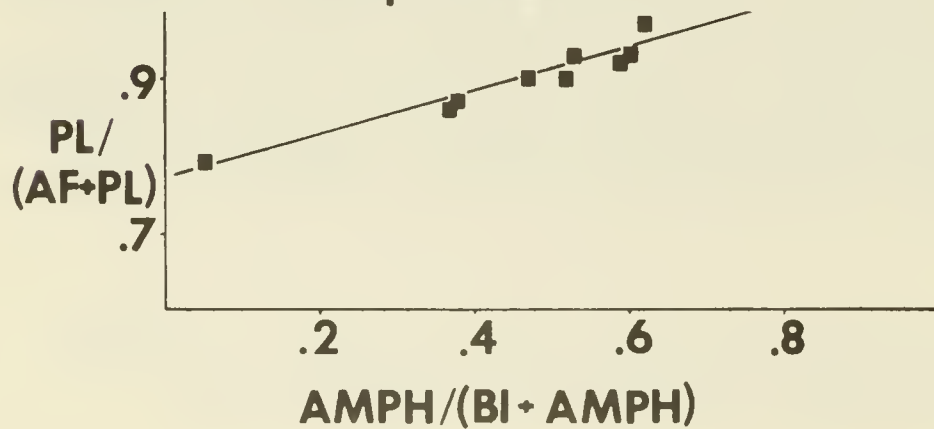
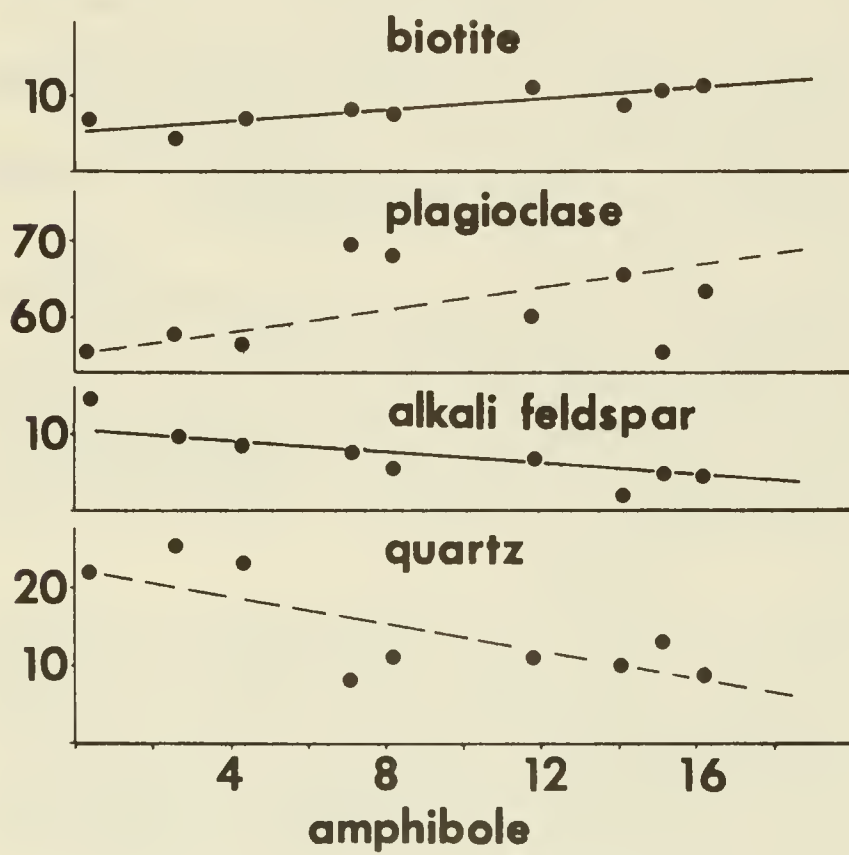
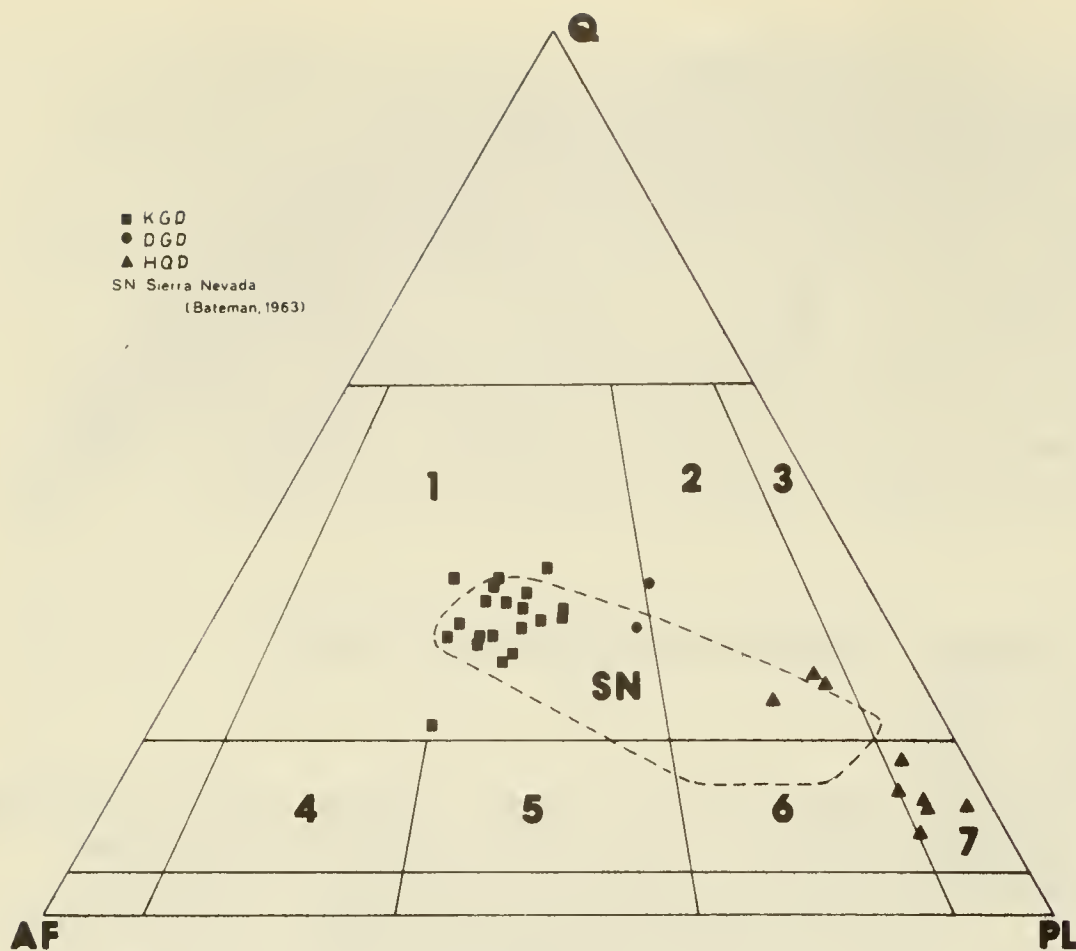


Figure 6. TOP: At top IUGS classification triangle for acidic plutonic rocks with plotted model analysis of rocks from the Katahdin Pluton.

BOTTOM: Mineral variation diagrams in Horserace Quartz Diorite.

Debsconeag Granodiorite

Debsconeag Granodiorite (DGD) is a medium grained amphibole-bearing biotite granodiorite found principally in a 80 km² large body in the central part of the Katahdin Pluton (Fig.4). Another occurrence of DGD is a small stock (1.5 km²) located in the southern part of the KP. The larger of the 2 stocks is intruded by dikes associated with the HQD.

DGD in hand specimen is similar to the Doubletop facies of the Katahdin granite but with a noticeably higher content of biotite and ever present traces of amphibole. Modal analyses are plotted in Fig.6. Typically, DGD contains alkali feldspar (21%), cyclically zoned plagioclases: AN35 to AN18 (37%), quartz (31%), biotite (9%), hornblende (1%), and accessory apatite, zircon, allanite and opaques.

Major Element Geochemistry

Major element analyses of representative samples of each of the rock types are plotted on an AFM (wt%) diagram and their normative compositions are plotted on an AB-Q-OR diagram (Fig.7). From these diagrams a clear distinction exists between the more reducing TR-KG sequence and the more oxidizing HQD-DGD series. AFM plots are compared to the Skaergaard trend (reducing trend) and to the Lower California Batholith trend (oxidizing trend). The trend shown for Lower California Batholith is also typical of calc-alkaline magmatism associated with destructive plate margins (high alumina basalt-andesite-dacite association). From major element geochemistry, it is also deduced that the Traveler Rhyolite at its origin was at 8 kb total pressure, 4 kb water pressure and in equilibrium with two feldspars and quartz. The low oxygen fugacity of the TR magma was presumably maintained through equilibrium with graphite, which is a common component of Appalachian eugeosynclinal sediments.

Trace Element Geochemistry

Fig.8 shows chondrite normalized REE abundances of all the rock types within the Katahdin Pluton. The REE, as well as other trace elements, clearly indicate that the two magmatic sequences are of different origin. Modeling of trace element abundances show that the Katahdin Granite can be derived from the Traveler Rhyolite magma if 15-25% phenocrysts are removed from the TR magma by fractionation. Debsconeag Granodiorite is best explained by mixing of approximately 40% Traveler magma and 60% Horserace magma. Trace elements also suggest that the TR magma could have been produced by 20% partial melting of eugeosynclinal sediments. Trace element and major element abundances of the Horserace Quartz Diorite are consistent with abundances of calc-alkaline associations of continental margins. If this inference is correct then the HQD magma may represent magma derived along a subducting oceanic crust.

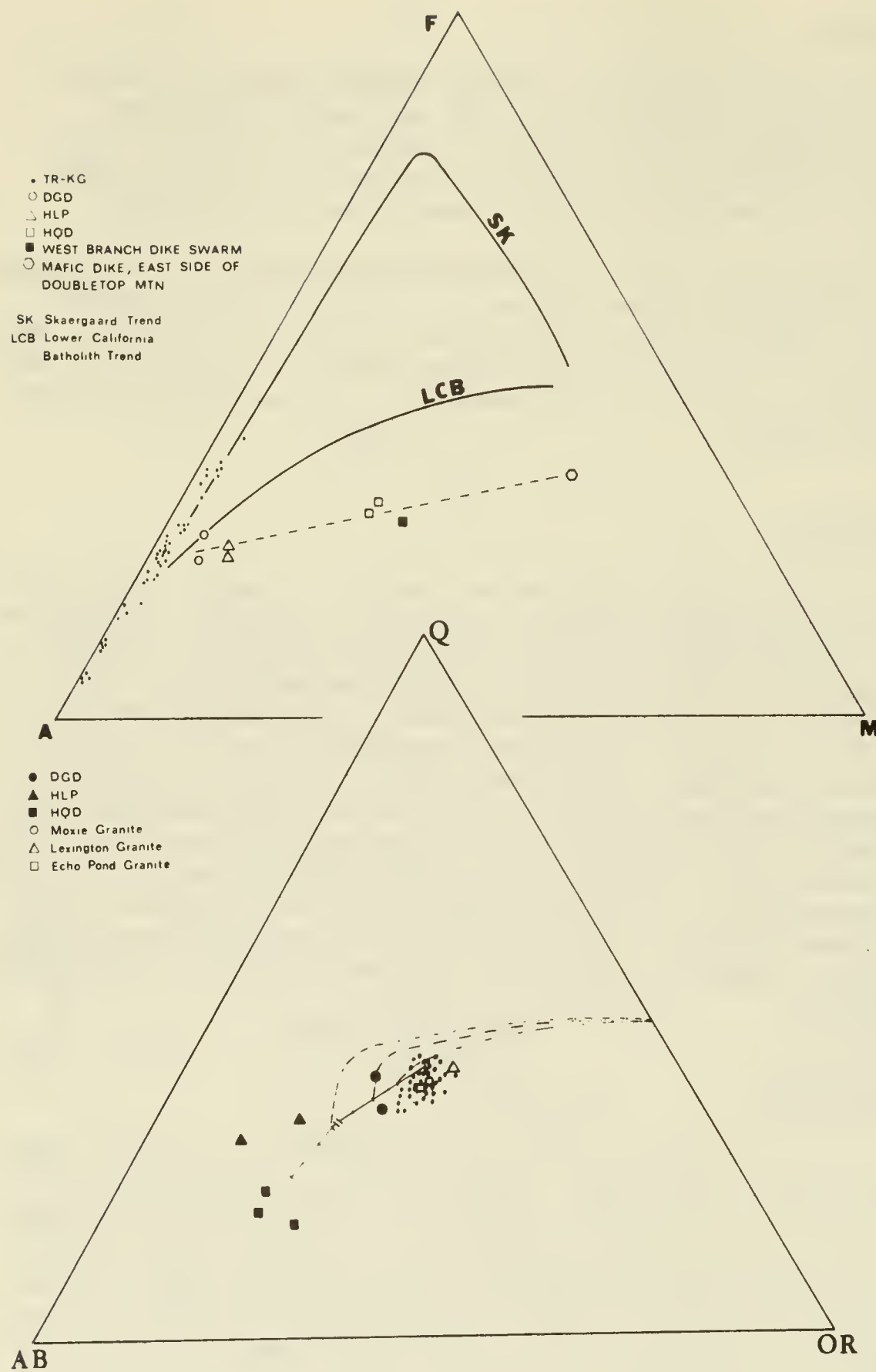


Figure 7. TOP: AFM (wt %) plot of chemical analysis for rocks within the Katahdin Pluton.
 BOTTOM: AB-Q-OR plot of normative compositions of Katahdin Pluton.

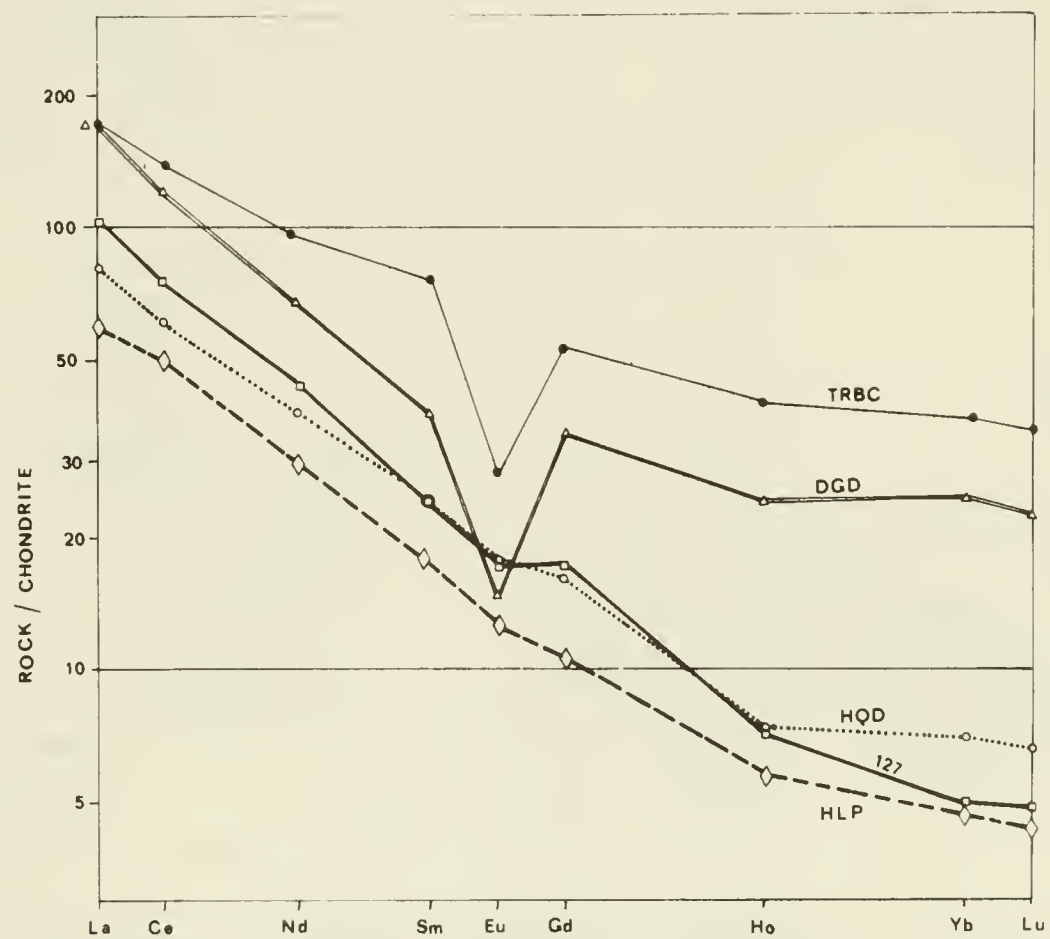
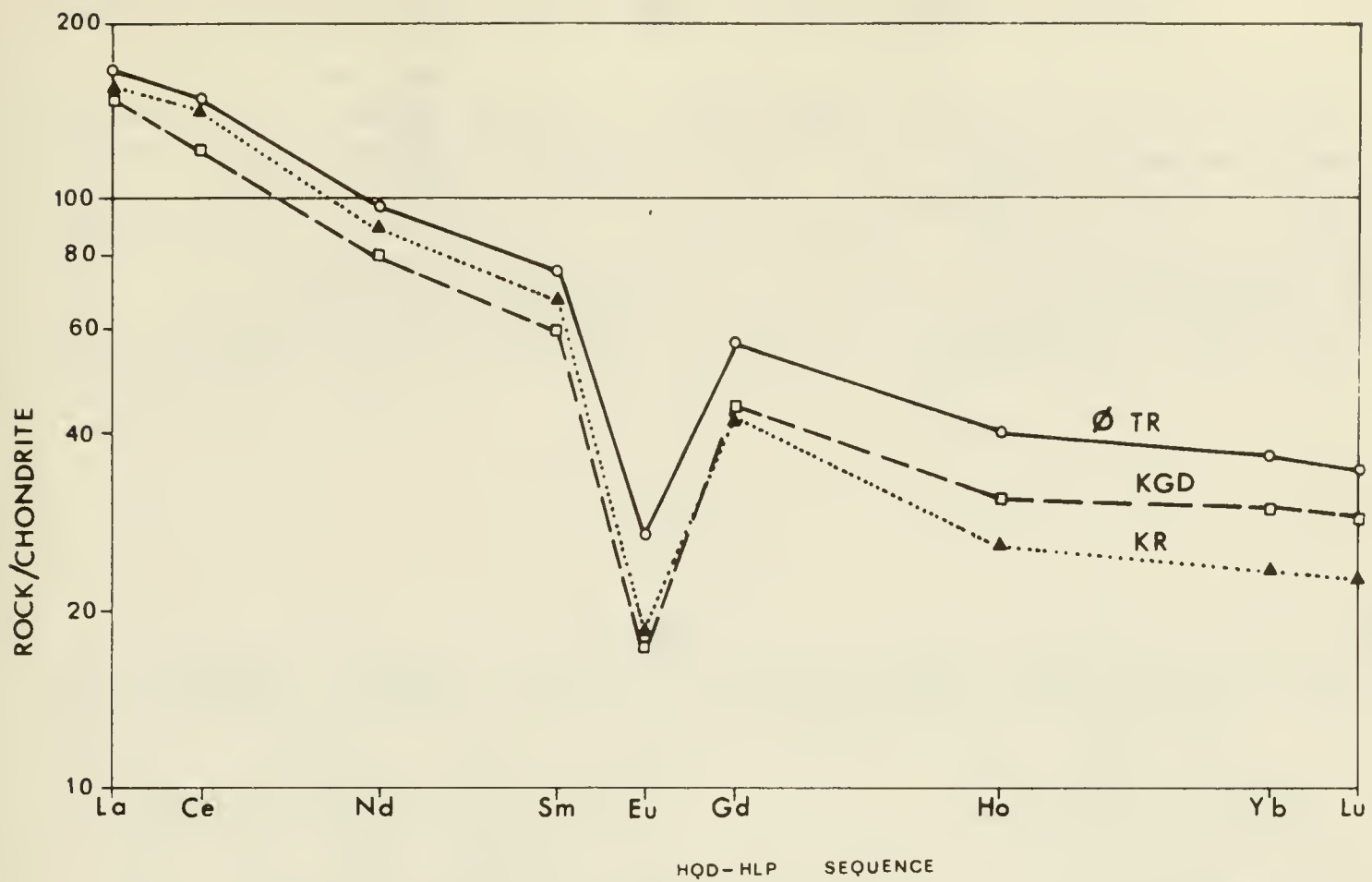


Figure 8. Chondrite normalized REE abundances for rocks of Katahdin Pluton. Sample 127 is from the dike associated with the HQD intrusion. KR-Kineo Rhyolite.

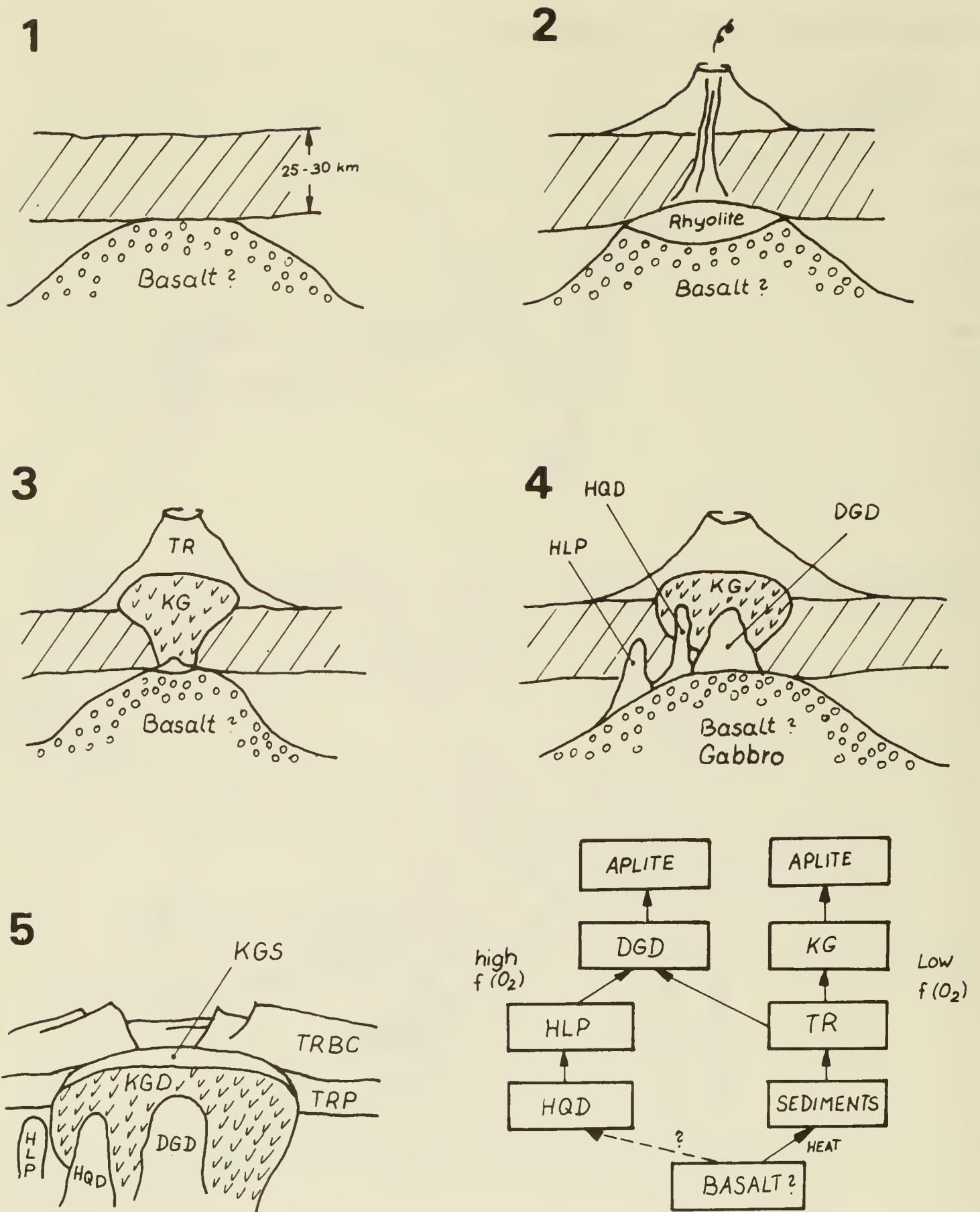


Figure 9. Schematic diagram of origin of Traveler Rhyolite - Katahdin Pluton Igneous Complex.

Origin of Traveler Rhyolite-Katahdin Pluton Igneous Complex

Fig.9 schematically illustrates the possible time sequence for the origin of the igneous complex. Mafic magma of mantle origin (possibly derived along a subducting plate) raises the temperature near the base of the crust to initiate partial melting. The resulting melt of rhyolitic composition is then intruded to shallow levels. From here it is rapidly moved to the surface through violent ash-flow eruptions. The residual melt fractionates and on crystallization forms the Katahdin Granite which is in turn intruded by smaller intrusions of quartz diorite (HQD) and granodiorite (DGD).

Statement

Since the exact route of the trip is somewhat weather-dependent, the trip log will be distributed at the first assembly place: Togue Pond Campground (outside southern entrance to Baxter State Park).

ALPINE GLACIATION OF MT. KATAHDIN

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Introduction

Mt. Katahdin is the highest mountain in Maine (1601m) and, with a local relief of about 1450 meters, is one of the largest mountains east of the Rocky Mountains. The mountain is composed of quartz monzonite (Katahdin granite informally) and is part of a large Devonian batholith that intrudes Lower and Middle Paleozoic sedimentary and volcanic rocks. The relief of the mountain can be explained in part by the greater amount of erosion of the sedimentary rocks compared with that of the more durable plutonic rock. However, the Katahdin pluton underlies much of the lowland south and east of the massif, so differential erosion can not completely explain the relief of the mountain above the surrounding countryside.

Erratics found by Tarr (1900) and Antevs (1932) near the summit of Mt. Katahdin and by Caldwell (1972) on other mountains in the region support the view that Mt. Katahdin was covered by continental ice sometime in the Pleistocene. There is no direct evidence that the highest elevations were covered by Late Wisconsin ice, although recent work by Davis (1978) suggests that they were.

Features Formed by Alpine Glaciation

Cirques

In the Mt. Katahdin area cirques occur only on those mountains underlain by Katahdin quartz monzonite. Most of the cirques (6) were formed on the massif which includes Mt. Katahdin and the other high peaks which lie above timberline. The 3 largest cirques are on the east side of the mountain and have headwall heights which range from 720 meters to about 100 meters. These 3 great cirques have flat to concave floors and steep headwalls composed largely of bedrock. Postglacial rockfall and avalanche debris does mask the lower slopes of the cirque headwalls and sidewalls. The total aspect of these cirques is one of remarkable freshness, especially when compared with other cirques in northeastern United States assumed to be occupied by glaciers in the Late Wisconsin (Craft, 1979).

Aretes

The three east-facing cirques are separated by aretes. The most typical arete is Hamlin Ridge, which separates North Basin cirque from Great Basin cirque (Figure 1). The arete which separates Great Basin cirque and South Basin cirque, Cathedral Ridge, has been shortened and



Figure 1. Oblique aerial view of Mt. Katahdin looking west. Numbers locate field trip stops which are discussed in text.

lowered by glacial erosion and mass wasting. The most spectacular serrate mountain crest is the Knife Edge but it may not technically be an arete because there is no cirque on its South side. However, the long narrow saw tooth ridge crest and the over 2000 foot (720m) drop into South Basin more than make up for this deficiency.

Moraines

The aspect of alpine glaciation on Mt. Katahdin about which there is the greatest controversy concerns moraines found within and down the mountain from the three largest cirques. Tarr (1900), Antevs (1932) and Caldwell (1965, 1972) identified moraines in each of these cirques. In addition these authors believed the large Basin Ponds moraine was a medial moraine formed between the combined alpine glaciers from the three cirques and the still active tongue of the Laurentide ice sheet. The common interpretation was that the alpine glaciers which formed these moraines were both contemporaneous with (at the Basin Ponds Moraine), and postdated (at the moraines within the cirques), the Late Wisconsin ice sheet. Davis (1978) believes the Laurentide ice sheet covered Mt. Katahdin during the Late Wisconsin but that no alpine glaciers postdated the ice sheet glaciation. Davis does not believe that there are moraines within the large cirques and interprets the Basin Ponds Moraine to be a lateral rather than a medial moraine.

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Road Log

Mileage

- 0 Togue Pond Camps. Starting time is 6:30 A.M. There is a fee of \$5.00 per car without State of Maine license to enter Park so double up in cars as much as possible. We might try to pay for permits the previous night to save time.
- 0.3 Enter Baxter State Park. Road is built on large esker.
- 0.4 Bear right at fork toward Roaring Brook campsite.
- 1.0 Rat Pond on left. If day is clear, the caravan will stop 10 minutes to allow photographs of Mt. Katahdin skyline.
- 1.8 Rum Brook.
- 5.0 Windey Pitch. Sandy washed drift and sandy till in which in 1966 there were sand-filled ice wedge structures. We can stop here on return trip. This ridge has no bedrock exposures and is believed to be an end moraine.
- 6.8 Avalanche Brook and Avalanche Field.
- 8.4 Roaring Brook Campsite. Bear left and park in designated area.

Trail Log

- 0 Trail log from Roaring Brook Campsite. Approximate mileage and hiking times (does not include time at stops) are given.
- 1.5
(45 min.) Stop 1. Halfway Rock (half way between Roaring Brook and Chimney Pond Campsites). The sandy drift exposed on trail and on small ridge to the right of trail has noticeable content of erratic fragments. Figure 1 shows location of this and other stops which are identified by numbers on the photograph.
- 1.9
(1 hr.) Stop 2. The Basin Ponds are dammed by the Basin Ponds Moraine (Figure 1, locality 2). The moraine is composed of about 99% Katahdin quartz monzonite and is inferred to have been deposited by ice issuing from the three great cirques to the west. The moraine has also been interpreted as a lateral moraine of the Late Wisconsin Laurentide ice sheet.
- 2.3
(1 hr.
15 min.) North Basin cutoff.

3.0
(2 hrs.)

Stop 3. Blueberry Knoll and North Basin cirque. Blueberry Knoll is at the mouth of what is probably the best preserved cirque in New England (Figure 1, locality 3). From Blueberry Knoll the Basin Ponds Moraine and associated hummocky topography to the east are clearly visible.

The evidence bearing upon the origin of Blueberry Knoll and the age of the last Alpine glaciation in the North Basin cirque will be reviewed. If we appear to be making good time, we should hike into the cirque to view the topography on the cirque floor at close range.

3.2
(2 hrs.
15 min.)

Hamlin Ridge trail. If the weather (and the leader) is still holding up, we will make the somewhat arduous climb up the Hamlin Ridge arete, in order to better view the features under discussion.

3.9
(4 hrs.)

Stop 4. Hamlin Ridge. In North Basin near the head of the cirque, but separated from the headwall by a depression some 60 meters wide and 6 to 12 meters deep, is a well defined accumulation of boulders. The feature resembles a protalus rampart, formed by rockfall rolling over permanent snow banks.

4.0
(4 hrs.
15 min.)

Stop 5. Hamlin Peak. Imposing and instructive views of North Basin, Hamlin Ridge, and Basin Ponds Moraine to the east, South Basin, the Knife Edge, and the Table Land toward the south. To the west, on the other side of the large basin called The Klondike, is a range of mountains with an unusual pattern in the forests. This pattern is produced by bands of dead but mostly standing trees. Although these bands may have something to do with the wind, they cannot be properly called blowdown or windthrow, as they commonly are.

4.3
(4 hrs.
30 min.)

Stop 6. Howe Peaks trail. View down the length of North Basin cirque.

4.5

Caribou Springs. Caribou were common on the mountain during the 18th and 19th centuries but were all killed off by 1900. In the 1960's there was an abortive attempt to repatriate the Caribou when several drugged, pregnant cows were lifted by helicopter to near this spot. After they came to, none were ever seen again, as far as I know.

5.2
(5 hrs.
15 min.)

Stop 7. The Saddle. Here we will separate the men (who will go down the mountain) from the boys (who may want to climb to the summit). The summit is a hard mile from here, our cars somewhat over 5 miles away. Those who accompany the leader down the saddle slide be very careful of loose rocks.

- 6.5 Chimney Pond. Spectacular view of the headwall of South
(6 hrs.) Basin beyond the rangers camp.
- 10.0 Roaring Brook and end of trip.
(7 hrs.
30 min.)

BEDROCK GEOLOGY OF THE SHIN POND-TRAVELER MOUNTAIN REGION

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Introduction

Low-grade metamorphism and the presence of fossils make the geology of northeastern Maine a key to understanding the more metamorphosed rocks of southern New England. The Shin Pond-Traveler Mountain region contains many of the critical elements of that key.

The region affords the longest and most complete, stratigraphic section in the State; it contains many fossil localities including an unusual occurrence of well-preserved Early Devonian terrestrial plants (see Trip B5), and it contains one of the largest masses of felsic volcanic rocks in the United States. Because facies change abruptly within short distances and because many critical relations are exposed in inaccessible places, the features to be seen along the routes of the trips are only random and incomplete samples of the information upon which the understanding of the geology of this area is based.

The geology of the quadrangles to be visited was mapped at a scale of 1:62,500. Neuman (1967) mapped the Shin Pond and Stacyville quadrangles for the U.S. Geological Survey; he has been especially interested in Ordovician stratigraphy and paleontology. Rankin (1961) mapped the Traveler Mountain quadrangle, with special emphasis on the volcanic rocks of Traveler Mountain, for a dissertation at Harvard University; he was supported in part by the Maine Geological Survey. The Island Falls quadrangle, east of the Shin Pond quadrangle, was mapped by E.B. Ekren and F.C. Frischknecht (1967) of the U.S. Geological Survey, using electromagnetic equipment to supplement surface observations. In addition to these reports, more than a dozen papers by these geologists on one or another aspect of their work have been published. Further, all but the mountainous area is covered by aeromagnetic maps of the U.S. Geological Survey.

We wish to acknowledge the essential role of Arthur J. Boucot, now of Oregon State University, in identifying and interpreting the many Silurian and Early Devonian brachiopods, and thus in establishing the relative ages of many of the units mapped. Graptolites in considerably fewer numbers were identified by W.B.N. Berry (University of California, Berkeley) whose assistance we also gratefully acknowledge.

The 1966 New England Intercollegiate Geological Conference in the Mount Katahdin Region (Caldwell, ed., 1966) included field trips led by Neuman and Rankin through much of the same area discussed here. Although the relevant sections of the 1966 guidebook are largely repeated here, information derived from additional work by Neuman, Rankin, and others has been added, and the itineraries of the trips have been modified to fit the time limitations of the 1980 program.

Major tectonic features

The major structures of the region are the large anticline that extends northeastward from the Stacyville quadrangle across the southern half of the Shin Pond quadrangle (the southwestern end of the Weeksboro-Lunksoos Lake anticline of Pavlides and others, 1964) and the complementary synclines to the northwest and southeast (fig. 1). Lower Cambrian(?) and Lower Ordovician rocks are exposed in the core of the anticlinorium. On the northwest flank of the anticline are Upper Ordovician rocks and a Silurian sequence of distinctive calcareous sedimentary rocks, conglomerate, and volcanic rocks that are overlain by Lower Devonian siltstone, sandstone, the volcanic rocks of Traveler Mountain, and the overlying sedimentary rocks that were derived from them. On the southeast flank of the anticline, by contrast, the Silurian rocks are largely a monotonous assemblage of slate, siltstone, and fine-grained sandstone, without volcanic rocks, and with little limestone or conglomerate; in this region no sedimentary rocks of Devonian age have been identified.

Most of the rocks are deformed and metamorphosed to the chlorite grade of regional metamorphism. Metamorphism and deformation are least in the Traveler Rhyolite and the overlying Trout Valley Formation, the latter being remarkably little disturbed.

The rocks of the Lower Cambrian(?) Grand Pitch Formation are intricately folded and faulted and are more deformed than those of overlying formations. Argillaceous rocks of the Grand Pitch possess a well-developed cleavage, and sandstones are commonly sheared. In many places cleavage is folded; in some of the cleavage folds the earlier cleavage is cut by a second one. Argillaceous rocks and interbedded sandstone at the base of the overlying Ordovician Shin Brook Formation at a few places are not as complexly deformed as those of the Grand Pitch. At other places the lower part of the Shin Brook contains conglomerate composed of fragments of slate and quartzite almost certainly derived from the Grand Pitch.

Both the deformation contrast and the clasts of the conglomerate indicate that a tectonic event separated the deposition of these formations. Deformation contrasts between rocks that may be correlative with the Grand Pitch and overlying Ordovician rocks have been described elsewhere in the northern Appalachians (Cooke, 1955; Riordon, 1957; Larrabee and others, 1965, p. E-8). Such a contrast through this large an area suggests tectonic activity of regional extent at some time between the Early Cambrian and the Early Ordovician; the term Penobscot disturbance was coined for this event (Neuman and Rankin in Caldwell (ed.), 1966, p. 9).

The effect of the Taconic orogeny in this area is indicated by several features. For example, the absence of Ordovician rocks beneath the Silurian in most places along the northwest flank of the Weeksboro-Lunksoos Lake anticline might be attributed to Taconic uplift and erosion. This event may also be responsible for the apparent wedge-out of

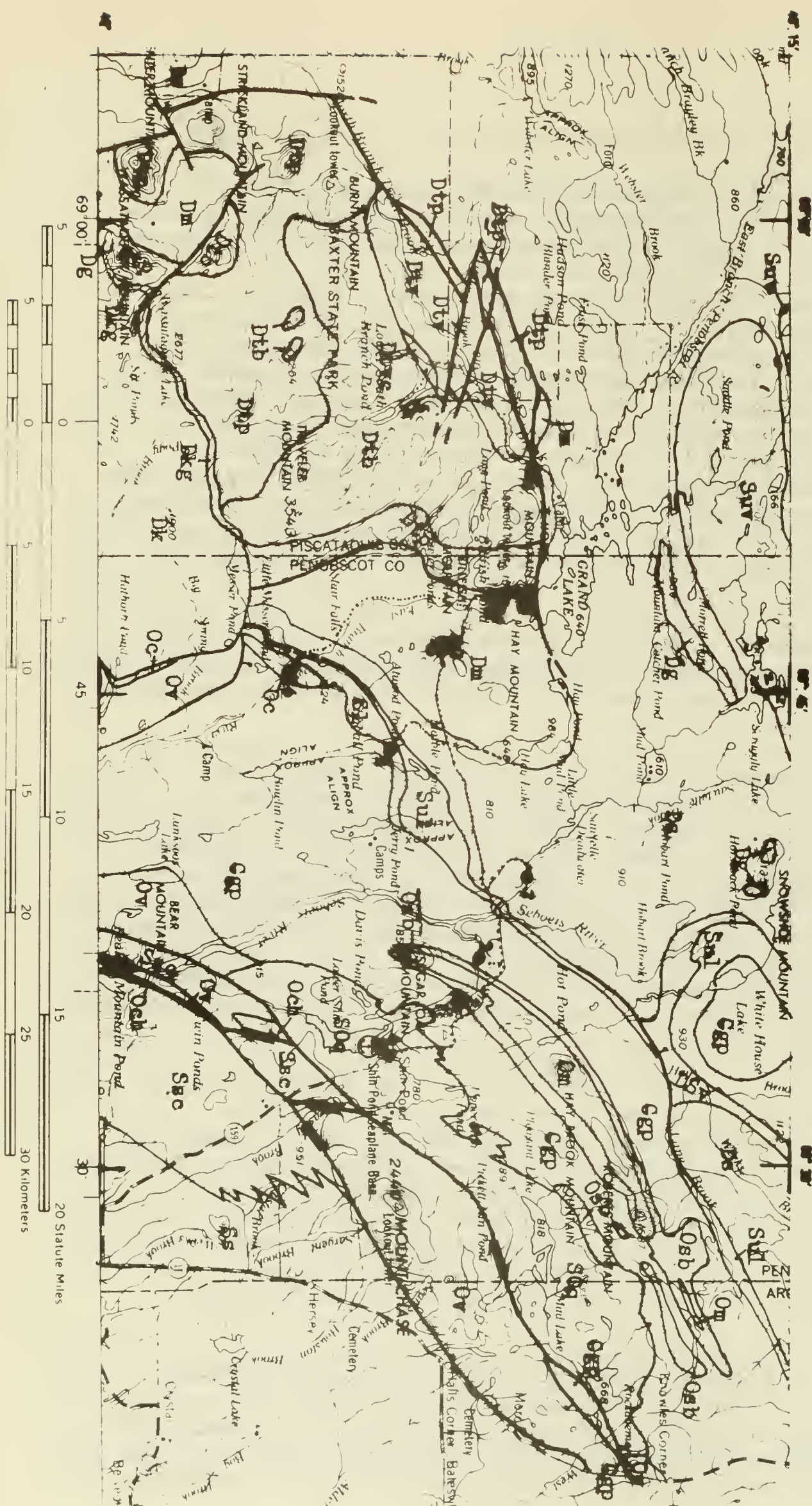
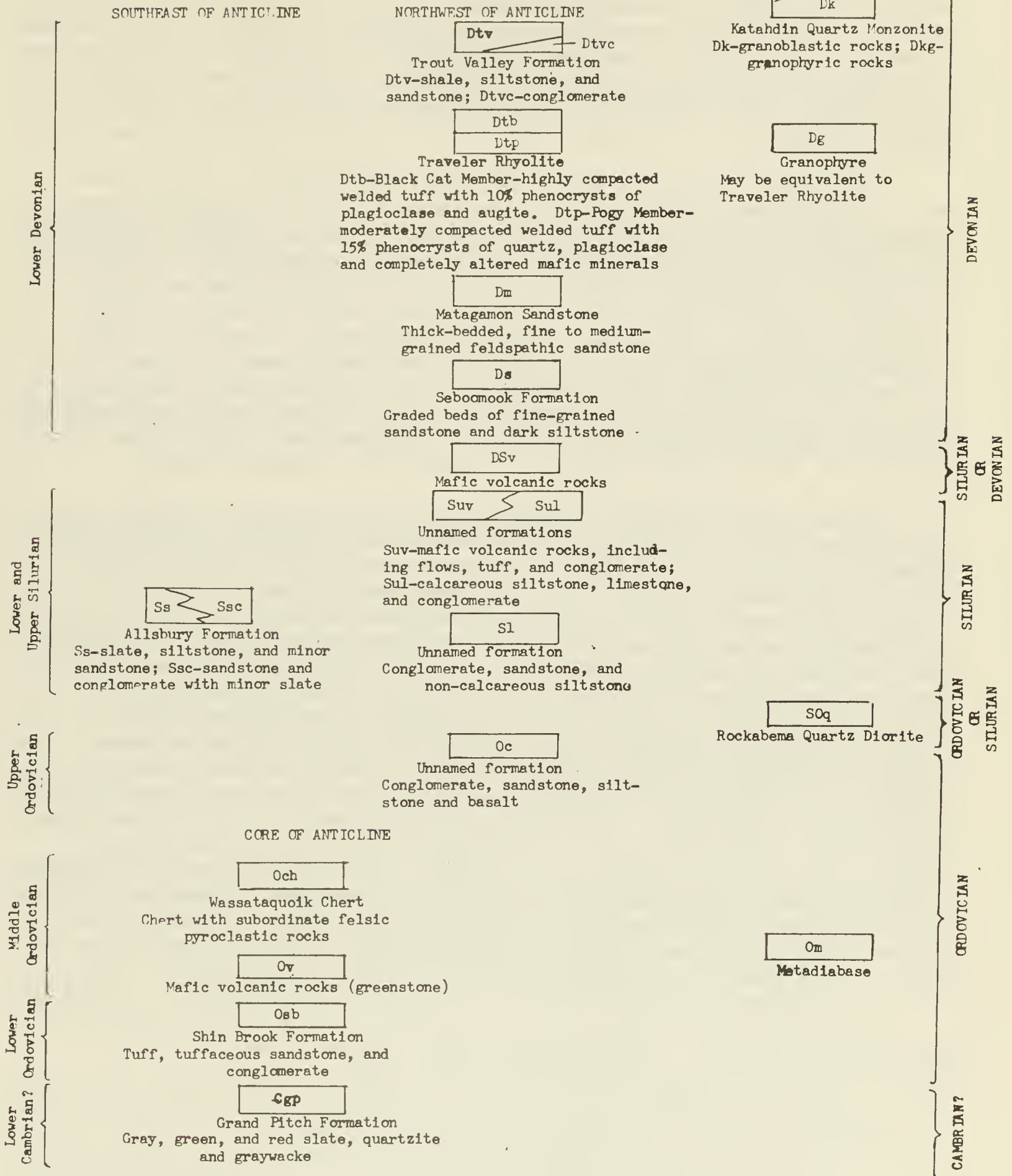


Figure 1. Geologic map of Traveler Mountain, Shin Pond, and part of the Island Falls quadrangles, and structure section approximately parallel to route of excursions; generalized from 1:62,500 scale maps cited in text.

EXPLANATION

SEDIMENTARY AND EXTRUSIVE ROCKS

INTRUSIVE ROCKS



Ordovician rocks at the southwestern end of this outcrop belt. The contrasting facies of contemporaneously deposited Silurian rocks on opposite sides of the Weeksboro-Lunksoos Lake anticline indicate that an ancestral form of this anticline developed during the Taconic and remained to separate the Silurian basins of deposition. Fragments of the Rockabema Quartz Diorite in Lower Silurian conglomerate on the southeast flank of the anticline were probably locally derived and indicate the minimum age of that intrusive.

The Acadian orogeny was the last major deformation to affect the area. Through most of the region, Acadian structure is characterized by nearly vertical, well-developed slaty cleavage and shear surfaces; folds are the dominant major structures, but there are significant contrasts in the style of folding on opposite sides of the Weeksboro-Lunksoos Lake anticline, and faults are important features in some places. On the southeast flank of the anticline most beds as well as cleavage stand nearly vertical; axes of minor folds and bedding-cleavage intersections are generally vertical. By contrast, on the northwest flank, bedding over wide areas dips moderately, and major as well as minor folds have moderate plunges. Curiously, over a considerable area east of Traveler Mountain, minor folds plunge northeast whereas major folds plunge southwest.

The age of the Acadian orogeny relative to the age of the Traveler Rhyolite and Trout Valley Formation poses some difficult questions. The Katahdin Quartz Monzonite lacks a tectonic fabric and clearly intrudes folded Lower Devonian rocks; it is a post-orogenic pluton. The structurally competent Matagamon Sandstone and Traveler Rhyolite are part of these folds, but these competent rocks show the effects of deformation less than the underlying rocks. The ash flow of the Pogy Member of the Traveler Rhyolite overrode unconsolidated sediments of Matagamon as evidenced by sandstone dikes in the basal Pogy flows and possible channeling of the Matagamon by the ash flows (Rankin, this guidebook). Pebbles of felsite in the upper few meters of the Matagamon provide further evidence that volcanism began before the end of the deposition of the Matagamon Sandstone. On the other hand, both Hon (this guidebook) and Rankin (this guidebook) conclude that the Katahdin Quartz Monzonite is a subvolcanic pluton of the Traveler caldera. Thus, there can be no great interval of time between the deposition of the Matagamon Sandstone, folding by the Acadian orogeny, and the intrusion of the post-orogenic Katahdin pluton. These observations lend further support to the suggestion by Naylor (1971) that the Acadian orogeny was a short-lived event.

The Trout Valley Formation is so little deformed that it may postdate the Acadian orogeny. The age of the Trout Valley, relative to the intrusion of the Katahdin Quartz Monzonite, is not known.

Stratigraphy

CORE OF ANTICLINE

Grand Pitch Formation (Neuman, 1962): Gray, green, and red slate and siltstone and about equal amounts of vitreous quartzite and lesser amounts

of graywacke and tuff. Contains the trace fossil Oldhamia smithi Ruedemann in red slate at several places along the East Branch of the Penobscot River. Similar rocks of the Nassau Formation in New York that have no fossils other than Oldhamia are considered to be of late Proterozoic age (Fisher, 1977); Oldhamia also occurs with Early Cambrian body fossils in the Weymouth Formation in Massachusetts (Howell, 1922), and with late Early to early Middle Cambrian acritarchs in the Bray Group of southeastern Ireland (Bruck, Potter, and Downie, 1974, p. 80). Minimum thickness, 1500 m (5,000 ft).

Shin Brook Formation (Neuman, 1964): Tuff, tuffaceous sandstone and conglomerate, breccia, and flows. Tuff, the most common rock, is massive, greenish-gray, and porphyritic; contains saussuritized stubby to anhedral plagioclase phenocrysts as much as 2 mm in cross section; it is of intermediate composition, in the andesite-dacite range. Fossils, mostly brachiopods, and fewer trilobites, bryozoans, gastropods, and sponges, occur in the sandstone and tuff at different levels from place to place. Paleontological studies of these and related fossils from New Brunswick, Newfoundland, and Wales indicate a late Early Ordovician age (e.g., Neuman, 1964, 1976; Neuman and Bates, 1978; G.S. Nowlan, written commun., 1979). Thickness variable, 100 to 750 m (300 to 2,500 ft).

Ordovician mafic volcanic rocks (greenstone): Largely massive, dark greenish-gray, locally pillow lava and flow breccia. Petrology summarized from Hynes (1976, p. 1216): Some rocks are highly porphyritic and contain both feldspar and pyroxene phenocrysts in fine fluidal groundmass. Alteration of some feldspars varies from core to margin suggesting that the feldspars originally were zoned. Some pyroxene phenocrysts have good oscillatory zoning. Pyroxene commonly less than 20 percent of mode. These observations indicate that the rocks are probably meta-andesites. Many coarse-grained rocks have ophitic texture; fine-grained vesicular rocks that have almost 40 percent pyroxene were probably originally basalts. The presence of both basalts and andesites is supported by bulk chemistry which ranges from 50 to 60 wt percent silica. Thickness where present, 300 to 750 m (1,000 to 2,500 ft).

Wassataquoik Chert: Chert with subordinate felsic and mafic pyroclastic rocks. Thin-bedded, medium- to dark-gray, greenish-gray, and red chert; tuff and tuff breccia interbedded locally. Siliceous shale interbeds contain graptolites of the Climacograptus bicornis and Orthograptus truncatus var. intermedius Zones, and conodonts and inarticulate brachiopods. Estimated thickness, 100 to 450 m (300 to 1,500 ft).

NORTHWEST FLANK OF ANTICLINE

Ordovician conglomerate, sandstone, siltstone and basalt: Polymict boulder to pebble conglomerate containing fragments of volcanic rocks, slate, quartzite, and quartz pebbles, with interbedded siltstone, ankeritic near the top; basalt at the base and near middle are dark greenish-gray, very fine-grained, with phenocrysts of plagioclase, pyroxene, and olivine; some pillow structures (Rankin, 1961, p. 41). Exposed principally along the East Branch of the Penobscot River

(including Haskell Rock Pitch), presumably overlain by Lower Silurian conglomerate; wedges out northeastward. Contains brachiopods, trilobites, and corals of Late Ordovician (Ashgill) age. Maximum thickness about 1,200 m (3,500 ft), but wedges out abruptly.

Lower Silurian conglomerate, sandstone, and siltstone: Thick-bedded polymict quartzose pebble conglomerate, micaceous sandstone, and gray and red siltstone and slate; wedges out northeastward. Conglomerate contains large thick-shelled brachiopods, such as Pentamerus sp. and Stricklandia lens ultima Williams. As much as 250 m (800 ft) thick.

Upper Silurian calcareous siltstone, limestone, and conglomerate: Light-gray calcareous siltstone and fine-grained sandstone containing thin beds and lenses of silty limestone; includes some reefal limestone at Marble Pond and elsewhere, and coarser grained sandstone and conglomerate in the northeast corner of the Shin Pond quadrangle. Fossils, especially brachiopods, corals, and stromatoporoids locally abundant. Some assemblages dated as Early or Late Silurian (late Llandoveryan or Wenlockian) age; others are more certainly Late Silurian (Wenlock or early Ludlow) age. Probable minimum thickness, 150 m (500 ft).

Upper Silurian mafic volcanic rocks (apparently a thick volcanic equivalent of the calcareous siltstone sequence described above): Massive metamorphosed mafic volcanic rocks including pyroclastics, interlayered with green tuffaceous slate and siltstone, conglomerate with red and green matrix, and muddy sandstone; also minor amounts of reefal limestone, some containing basaltic clasts. Scattered fossils in green tuffaceous slate, green matrix conglomerate, reefal limestone and debris derived therefrom; some assemblages dated as Late Silurian (early Ludlovian), others dated no more precisely than Silurian or Devonian. Some pre-Silurian rocks may be included. Thickness a thousand meters or more (several thousand feet).

Devonian or Silurian mafic volcanic rocks: Tuff, breccia with scoriaceous fragments, and probably some flows. Possibly the same as Upper Silurian volcanic unit, but lacks fossils.

Seboomook Formation (Boucot, 1961, p. 169): Graded beds of fine-grained, cross-bedded sandstone, dark-gray siltstone, slate, and a few thick beds of fine-grained feldspathic sandstone like that of the Matagamon Sandstone. One exposure of gray sandy siltstone at the base contains a few Early Devonian brachiopods. Primarily a submarine-slope and prodelta deposit according to Hall and Stanley (1973, p. 2101). Thickness variable: 1,200 m (4,000 ft) on East Branch of the Penobscot River.

Matagamon Sandstone (Rankin, 1965): Thick-bedded, fine- to medium-grained feldspathic sandstone and subordinate amounts of siltstone and slate like that of the Seboomook. Sandstone commonly well laminated and crossbedded; some displays scour-and-fill structure. Load casts of sandstone in siltstone ("pseudonodules") rare. The Matagamon is a sandstone facies of the Seboomook. Fossils scarce except in occasional shell beds where Early

Devonian (Becraft-Oriskany) brachiopods are abundant. Primarily delta-top and delta-front deposits of a westerly Prograded delta (Hall and Stanley, 1973, p. 2101). Thickness, 1,200 to 1,500 m (4,000 to 5,000 ft).

Traveler Rhyolite (Rankin, 1968): Flinty aphanitic rhyolite that breaks with a conchoidal fracture and contains 10 to 15 percent of small (1 to 3 mm) phenocrysts. Color ranges from light gray through various shades of greenish, greenish gray and bluish gray to nearly black. In general, the darkest rocks contain the least altered phenocrysts and the best preserved primary textures. Largely welded ash-flow tuff, minor breccia, and rare airfall tuff and sandstone shale. Columnar jointing characteristic of the welded tuff. Younger than the Matagamon Sandstone of Becraft-Oriskany age and older than the overlying Trout Valley Formation of late Early or Middle Devonian age. Youngest stratigraphic unit intruded by the Katahdin Quartz Monzonite.

Bottino and others (1966) determined a Rb-Sr whole-rock isochron of 360 m.y. \pm 10 m.y. ($\lambda_{\text{Rb}^{87}} = 1.39 \times 10^{-11} \text{ year}^{-1}$) for the Traveler Rhyolite. This is in conflict with its stratigraphic position below the Trout Valley Formation and with the 395 m.y. age of the Katahdin Quartz Monzonite.

The Traveler Rhyolite is composed of two members:

Pogy Member - Lower member. Moderately compacted welded ash-flow tuff containing about 15 percent phenocrysts of quartz, plagioclase and completely altered mafic minerals.

Estimated thickness, 900 m (3,000 ft).

Black Cat Member - Upper member. Highly compacted welded ash-flow tuff containing about 10 percent phenocrysts of plagioclase and augite.

Estimated thickness, 2,300 m (7,500 ft).

Trout Valley Formation (Dorf and Rankin, 1962): Light blue-gray to black shale, siltstone, sandstone, conglomerate, and minor amounts of sideritic sandstone and black sideritic ironstone. A massive conglomerate lentil, probably a deltaic deposit, is present at the base along South Branch Ponds Brook -- the route traversed by Field Trip B3. Although pebble and granule conglomerate is scattered throughout, conglomerate lenses are less common in the upper part; boulder and cobble conglomerate is largely restricted to the basal conglomerate lentil. No rock fragments other than felsite have been observed in the conglomerate.

Fossils include plants (in some places so abundant that the rock resembles a low-grade coal), ostracodes, estherids(?), and eurypterid scales. Well-preserved terrestrial plants dominated by Psilophyton indicated a late Early Devonian (Onesquethawan-late Coblenzian) age to Dorf and Rankin (1962), but possibly an early Middle Devonian age to Schopf (1964, p. D49) and Andrews and others (1977, p. 283). The relatively undeformed condition of the Trout Valley may be due to its post-tectonic age if it proves to be equivalent to the post-Acadian Middle Devonian Mapleton Sandstone of Aroostook County; on the other hand, it may be due to its shielded tectonic position above the thick competent Traveler Rhyolite. Exposed thickness about 450 m (1,500 ft).

SOUTHEAST FLANK OF ANTICLINE

Allsbury Formation (Ekren and Frischknecht, 1967): Sandstone conglomerate, and minor slate - feldspathic sandstone, polymict pebble and cobble conglomerate, and gray slate and siltstone. The coarser conglomerate, containing cobbles of porphyritic quartz diorite, like the Rockabema, greenstone, quartzite, and other rocks, occurs in the fault slices of the southeastern flank of the anticline; at one place interbedded sandstone yielded Early Silurian (late Llandoveryan) fossils. Estimated minimum thickness, 1,500 m (5,000 ft). Slate, siltstone, and minor sandstone - medium- to dark-gray, greenish-gray, and red slate and siltstone, and a few beds of fine- to medium-grained sandstone. Monograptid graptolites rare, including late Llandoveryan to early Ludlovian forms. Estimated thickness, about 3,000 m (10,000 ft).

INTRUSIVE ROCKS

Ordovician metadiabase: Gray and greenish-gray, fine- to coarse-grained metadiabase forming massive ledges. Forms as a sill above Shin Brook Formation.

Rockabema Quartz Diorite (Ekren and Frischknecht, 1967): Fine- to coarse-grained, gray to greenish-gray, sheared and altered porphyritic quartz diorite and granodiorite, characterized by phenocrysts of quartz and feldspar as much as half an inch in cross section. Potassic feldspar, some slightly perthitic, constitutes as much as one-third of the feldspar. Total feldspar somewhat more abundant than quartz. Chlorite and epidote pseudomorphic after biotite; calcite abundant in patches and veinlets. Locally contains abundant large xenoliths of greenstone and quartzite.

Devonian granophyre: Light-gray granophyre containing about 5 percent phenocrysts of quartz, plagioclase, and biotite. Plagioclase phenocrysts commonly in rosettes 2 to 3 mm in diameter. Groundmass granophyric or spherulitic.

Katahdin Quartz Monzonite (Neuman, 1967): Hypidiomorphic phase (=granoblastic phase of fig. 1) is massive medium gray, medium grained, and consists of two-thirds feldspar (about three-fifths perthite and two-fifths zoned plagioclase), one-third quartz, and 5 to 10 percent biotite. Where altered, potassic feldspar is pink, plagioclase is greenish, and chlorite replaces biotite. The quartz monzonite is porphyritic locally, and contains pink-weathering perthite phenocrysts 5 mm long in a groundmass somewhat finer grained elsewhere. The granophyric phase is vuggy, pink, and contains phenocrysts of biotite. Vugs contain epidote, tourmaline, quartz, and potassic feldspar. Border phase on the east is fine grained and contains abundant fragments of thermally altered and partially assimilated sedimentary rocks.

A preliminary Rb-Sr whole-rock isochron age of 395 m.y. ($\lambda_{\text{Rb}}^{87} = 1.39 \times 10^{-11} \text{ year}^{-1}$) for the Katahdin Quartz Monzonite was determined

Naylor and others (1974). This is a somewhat older age than the K-Ar ages obtained for the Katahdin east of Ripogenus Dam (356 m.y.) and a diorite stock at Nesowadnehunk Deadwater (361 m.y.) within the Katahdin batholith (Faul and others, 1963). The 395 m.y. figure is close to several other K-Ar ages reported by Faul and others (1963) for plutons in northern Maine and to the K-Ar age of 390 m.y. of a small pegmatite, presumably from the Katahdin batholith, in border-phase breccia east of Mt. Katahdin.

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TRIP B-3

THE TRAVELER RHYOLITE AND ITS DEVONIAN SETTING,
TRAVELER MOUNTAIN AREA, MAINE

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The stratigraphic and structural setting of the Traveler Rhyolite have been outlined by Neuman and Rankin in this guidebook. Before describing the field trip, I would like to discuss the Traveler Rhyolite in more detail.

The main body of the Traveler Rhyolite occupies a structurally depressed, roughly quadrilateral area about 13 by 19 km on the northwest limb of the Weeksboro-Lunksoos Lake anticline (see fig. 1 of Neuman and Rankin, this guidebook). The aggregate thickness of the volcanic pile within the depression is at least 3,200 m (10,500 ft). The main body of rhyolite is bounded by high-angle faults to the north and west and is intruded by the Katahdin Quartz Monzonite on the south. The depression is thought to be an ancient caldera called the Traveler caldera (Rankin, 1968).

The Traveler Rhyolite is typically a dark-greenish-gray or bluish-gray to nearly black aphanitic rock that breaks with a conchoidal fracture. A whitish weathering rind is common. The rhyolite is porphyritic and contains 3 to 30 percent (commonly 5 to 15 percent) small phenocrysts 1 to 3 mm across.

Many outcrops of the rhyolite offer little evidence about the type of volcanic activity that produced the rock. Thin sections help decipher some relevant features of these outcrops, but many rocks are too recrystallized to classify. Ash-flow tuff (both welded and nonwelded), tuff breccia, bedded air-fall(?) tuff and crystal tuff, lava, and tuffaceous sedimentary rock have been identified. Enough evidence is preserved, however, to establish that the great bulk of the Traveler Rhyolite consists of welded ash-flow sheets. Little sedimentary rock and no soil horizons have been identified within the volcanic pile, suggesting that the pile was erupted in a relatively short time. The dominance of welded tuff argues for subaerial eruptions. That the rhyolite welded ash-flow sheets overlie sedimentary rocks containing marine fossils (Seboomook Formation and Matagamon Sandstone) requires a change in the paleoenvironment from marine to terrestrial (Rankin, 1960).

Evidence that the bulk of the Traveler Rhyolite is welded ash-flow tuff comes from both outcrop and thin section. The features preserved include flattened pumice lumps, deformed (flattened) shards, columnar joints and massive units without obvious bedding. No glass remains in these Devonian rocks, and welding per se cannot be demonstrated, but flattening certainly can.

Collapsed pumice lumps, giving rise to a crude planar structure, are obvious in many outcrops but difficult to observe in others. Excellent

examples of rhyolite containing collapsed pumice lumps will be seen along South Branch Ponds Brook, in loose pebbles on the shore of Lower South Branch Pond, and, to a lesser extent, in the rhyolite of Horse Mountain. Some difference in the average size of the pumice lumps is apparent from one part of the volcanic pile to another, or even from outcrop to outcrop. Detailed mapping eventually may demonstrate that individual flow units are traceable, but so far, tracing of flow units has not been attempted.

Columnar joints are obvious in many outcrops and cliff exposures. They range in diameter from about 2 cm to 2 m; the larger ones are as much as a few tens of meters long. Spectacular thin columns, averaging about 8 cm in diameter and a few meters in length, are exposed along Dry Brook. Even smaller columns, some of which are bent, are found on the North Ridge of North Traveler Mountain. We will not have time to visit either locality, but we will see well-developed larger columns at Horse Mountain and along South Branch Ponds Brook.

Some mountain slopes have a terraced appearance, suggesting the erosion of a layered sequence. Commonly, the rises of these terraces are zones of well-developed columnar joints, whereas the treads are zones of indistinct columns or no columns. In a few sequences, the size of the pumice lumps changes from one terrace rise to another suggesting that the rises represent different flow units of a welded tuff sequence. More commonly, flow units and cooling units cannot be distinguished from one another or mapped (for terminology, see Smith, 1960).

The compaction foliation (eutaxitic texture) caused by the flattening and welding of pumice lumps and shards is commonly roughly perpendicular to the axes of the columns. Where this relationship holds true, the direction of compaction was roughly perpendicular to the cooling surface, presumably a quasi-horizontal surface, and the present attitude of the compaction foliation is a measure of subsequent deformation. Columns having axes that are not perpendicular to the compaction foliation may indicate that the cooling surface was not horizontal or that the compaction foliation was rotated by strain during deformation.

Tilted benches and measured attitudes of compaction foliation on Black Cat Mountain, and on North Ridge, Center Ridge, and Pinnacle Ridge of Traveler Mountain indicate that the axis of an open north-plunging anticline runs through the South Branch Ponds. This fold is readily apparent from the south end of Upper South Branch Pond on the trail up Center Ridge. Compaction foliation dips of 20° to 40° are common. Excellent columnar joints can be seen on the basal cliff of Center Ridge at the northeast end of Upper South Branch Pond.

The Traveler Rhyolite is divided into the basal Pogy Member and the overlying Black Cat Member. As explained below, the upper part of the Black Cat Member probably could be mapped as a third member.

The Pogy Member is characterized by moderately compacted welded ash-flow tuff containing about 15 percent phenocrysts, of which about one-third are quartz and most of the rest are zoned plagioclase. Optical studies

indicate that the average composition of the plagioclase is about An_{48} and zones range from An_{35} to An_{56} (Rankin, 1961). Sanidine (about Or_{67}) occurs with plagioclase and quartz in a few samples. Mafic silicate phenocrysts may be absent or may constitute as much as 10 percent of the phenocryst population. They are generally altered beyond recognition, but clinopyroxene was observed in one thin section, and the remnants of biotite phenocrysts were observed in a few thin sections. Garnet was seen in two samples. Lithic fragments, including rhyolite, diabase, sandstone, and shale, are present in most samples.

The Black Cat Member is characterized by highly compacted and welded ash-flow tuff containing about 10 percent phenocrysts. Typically, 75 percent of these are zoned plagioclase (An_{33} to An_{61} , bulk composition about Ab_{47}), 20 percent are augite (optically determined to be about $Wo_{33}En_{28}Fs_{42}$), and 5 percent are magnetite (Rankin, 1961). Quartz phenocrysts are typically not present but rarely constitute as much as 10 percent of the phenocrysts. Biotite phenocrysts coexist with augite in a number of samples, and rarely, hornblende is the only mafic silicate present. Fayalite coexisting with augite and biotite was observed in one thin section, and garnet phenocrysts were seen in two samples. Growth aggregates of phenocrysts, rare in the Pogy Member, are common in the Black Cat Member (fig. 1E).

An outlier of the Black Cat Member forms Soubunge Mountain about 14 km southwest of Strickland Mountain. The preserved thickness of rhyolite at Soubunge Mountain is about 250 m (800 ft), and it rests on a sandstone similar to the Matagamon (Andrew Griscom, written commun., 1966). In some of the Soubunge samples, hornblende phenocrysts are more abundant than augite.

The ash-flow tuff on the summit of Big Peaked Mountain (Traveler Mountain quadrangle), on Little Peaked Mountain, on Barrell Ridge, and along Dry Brook is currently included in the Black Cat Member; it may be a distinct unit at the top of the volcanic pile. In this tuff the pumice lumps tend to be smaller and less compacted and to form a smaller percentage of the rock than they do in the main part of the Black Cat Member. Phenocrysts make up about 5 percent of the rock and are thus less abundant than in the rest of the Black Cat. Plagioclase and augite are present. The rock is highly jointed; some jointing is almost a fracture cleavage.

The percentage of quartz phenocrysts is the only consistent difference observable in the field between the Pogy and Black Cat Members. The significance of the quartz phenocryst content was recognized after I completed most of the fieldwork. Limited field checking has shown that the subdivision is a valid one. It is, however, extremely difficult using a hand lens to identify small quartz phenocrysts that constitute 5 percent or less of the rock. For much of its length, the contact between the members is approximated between locations from which hand specimens had been collected previously.

From the evidence assembled, both the Pogy and Black Cat Members consist dominantly of welded ash-flow sheets. The phenocryst content of

the Pogy is somewhat greater than that of the Black Cat and differs significantly in that quartz is ubiquitous and sanidine may be present. The Pogy Member is typically more altered than the Black Cat and more commonly contains lithic clasts. The pumice lumps in the Black Cat are characteristically more compacted than those of the Pogy. Length-to-thickness ratios of 10:1 to 20:1 are typical for the collapsed pumice lumps in the Black Cat. Ratios as high as 60:1 have been observed. In the Pogy Member, on the other hand, ratios of 2:1 to 4:1 are more common.

As of this writing, the bulk composition of the Traveler Rhyolite has not been well sampled. The three analyses published by E.S.C. Smith (1930 and 1933), the two analyses by Jun Ito, reported by Rankin (1961), and one analysis by the U.S. Geological Survey (Rankin, unpublished data) do not indicate significant difference in bulk chemistry between the Pogy and Black Cat Members. This apparent uniformity in bulk chemistry is consistent with trace-element data for several samples of the rhyolite analyzed by Rudolph Hon (oral commun., 1975) as part of his Ph.D. thesis study on the Katahdin batholith. In August 1979, I collected a suite of 19 samples, including 3 from the Pogy Member, 12 from the main body of the Black Cat Member, and 4 from the upper part of the Black Cat Member. The major-element analyses of these samples are not yet available.

The silica content of the available analyses ranges from 71.62 to 72.54 weight percent (recalculated to 100 percent on a water-free basis) for two samples of the Pogy Member and from 70.50 to 75.06 percent for four samples of the Black Cat Member. The Differentiation Index of Thornton and Tuttle (1960) ranges from 82.6 to 84.5 for the two Pogy samples and from 79.9 to 88.7 for the four Black Cat samples. To a first approximation, the bulk compositions thus fall within the synthetic granite system. The normative constituents for both members plot in the vicinity of the minimum melting compositions in the steam-saturated system SiO_2 - KAlSi_3O_8 - $\text{NaAlSi}_3\text{O}_8$ (Tuttle and Bowen, 1958).

The model presented in 1968 (Rankin, 1968) to account for the observed differences between the Pogy and Black Cat Members still seems attractive. The arguments are summarized here from that paper. At the time of eruption, both quartz and feldspar were crystallizing from the magma that produced the Pogy Member, whereas liquids of the Black Cat Member were crystallizing only feldspar. The slightly higher phenocryst content of the Pogy further indicates that the Pogy was somewhat more crystallized at the time of the eruption. These observations are consistent with the hypothesis that the basal Pogy Member was erupted from the cooler upper part of the magma chamber and that the overlying Black Cat Member followed, probably relatively quickly, from deeper, hotter parts of the magma chamber.

Whether or not an ash flow will weld depends upon several variables, a major one of which is emplacement temperature (Smith, 1960, and Boyd, 1961). Calculations by Boyd (1961) show that a magma having a lower initial H_2O content will erupt an ash flow having a higher emplacement temperature than one erupted by a magma having a higher initial H_2O content. Other factors being equal, one would expect a drier magma to form a more highly compacted welded tuff than a wetter one. In fact, the

emplacement and welding temperatures of the Black Cat Member were high enough so that the tuff locally flowed after or during welding, as judged by the rotated phenocrysts (fig. 1F) and microscopic to mesoscopic flow folds.

At equilibrium, the H_2O content of a magma should increase upward in the magma chamber (Kennedy, 1955). If the Pogy Member originated from higher in the magma chamber than the Black Cat, it may have been wetter as well as cooler. This suggested gradient in water content of the magma chamber is also consistent with the phenocryst assemblage of the two members. Tuttle and Bowen (1958) showed that at the liquidus, a decrease in water content (decrease in steam saturation pressure) shifts the quartz-feldspar field boundary toward the SiO_2 corner of the $SiO_2 - KAlSi_3O_8 - NaAlSi_3O_8$ system. That is, a liquid in equilibrium with quartz and feldspar crystals at a higher steam saturation pressure might be in equilibrium with only feldspar crystals at a lower steam saturation pressure even though the composition of the liquid may be otherwise unchanged. This relationship was incorrectly stated in the 1966 NEIGC guidebook.

Corroborative evidence that the H_2O content of the Pogy was higher than that of the Black Cat comes from the widespread alteration of the mafic silicate phenocrysts in the Pogy as well as the more commonly altered plagioclase phenocrysts in the Pogy. This alteration may be deuteric, caused by abundant volatiles rising through the Pogy ash flows after emplacement. That the emplacement temperature of the Black Cat was higher than that of the Pogy is indicated by the greater degree of flattening of the Black Cat pumice lumps and the evidence for postemplacement flowage of the Black Cat. The existing evidence is consistent with the Pogy Member having had a lower magma temperature and emplacement temperature and a higher H_2O content (steam saturation pressure) than the Black Cat Member. These suggestions can be related to gradients in temperature and H_2O content in the magma chamber.

Partial results from the analyses of the suite of 19 samples collected in August 1979 indicate that the Traveler magma chamber was compositionally zoned in components other than H_2O . As stated above, the major-element chemistry for these samples is not available at this time, the trace-element analyses are only partially completed, and the Pogy Member is poorly sampled. The work is continuing. Quantitative analyses of trace elements show that compositional gradients are most obvious in the main body of the Black Cat Member but that more subtle gradients exist in the pile as a whole. The results appear to confirm that the upper part of the Black Cat Member on Big Peaked Mountain, Little Peaked Mountain, and Barrell Ridge is a different unit. In summary, Sr and Zr (by x-ray fluorescence), Nb (by spectrophotometry), and Li (by atomic absorption spectroscopy) decrease upward through the volcanic pile, and U (by delayed neutron activation analysis) increases upward in the pile. Semiquantitative emission spectrographic analysis indicated that the boron content of the Pogy Member is higher than that of the Black Cat Member.

Except for the outlier of the Black Cat Member on Soubunge Mountain, nothing is preserved today of the Traveler Rhyolite outside the proposed

Traveler caldera. If we make a conservative estimate for the average thickness of the rhyolite within the caldera of 1600 m, the volume of rhyolite within the structural depression is about 400 km^3 . A blanket of welded tuff 250 m thick (the thickness on Soubunge Mountain) surrounding the rim of the Traveler caldera to a distance as far as Soubunge Mountain would more than double the volume of material erupted. If the original volume of rhyolite was as much as 800 km^3 , the Traveler rhyolite represents one of the larger pyroclastic flow fields of the world (see Smith, 1960). If the Katahdin pluton, which has an area of about $1,350 \text{ km}^2$ (Griscom, 1976), represents the subvolcanic magma chamber as both Hon (this guidebook) and I think, then $400\text{--}800 \text{ km}^3$ of rhyolite is not unreasonable.

The Traveler Rhyolite is the northeasternmost and by far the largest of a discontinuous belt of similar rhyolitic volcanic and shallow granophyric intrusive rocks of Early Devonian age that extends about 160 km across north-central Maine. They are collectively called the Piscataquis volcanic belt (Rankin, 1968). Five major volcanic centers have been identified in the belt. Coeval mafic rocks, with the possible exception of the Moxie pluton, are conspicuously absent from the Piscataquis volcanic belt. Much of the rhyolite from these centers is peraluminous as evidenced by normative corundum in the available major-element analyses and garnet phenocrysts in the rocks. The Traveler Rhyolite appears to be the least peraluminous (that is, to have the lowest percentage of normative corundum) of the rhyolites, but garnet phenocrysts are present in a few samples of the Traveler. Garnet phenocrysts from the rhyolite of the Piscataquis volcanic belt range in composition from about 80 to 90 percent almandine; most of the remaining component is pyrope (Rankin, unpublished probe data).

I previously suggested that the Piscataquis volcanic belt was an island arc (Rankin, 1968). The observations that the volcanic center's overlies Lower Devonian marine sandstones, a shallower water facies than that of the surrounding Seboomook turbidites, and that many of the rhyolites must have been erupted subaerially are still valid. That is, the Early Devonian rhyolitic volcanism took place along a welt or ridge within a deeper water marine basin. That this volcanism was in any way related to the subduction of oceanic crust now seems unlikely.

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Itinerary, Trip B-3

Topographic quadrangle maps:

15-minute

Shin Pond

Traveler Mountain

2-degree

Presque Isle

Assemble at site of the former Shin Pond House, Maine Route 159, 9.5 miles northwest of Patten and 0.2 miles west of bridge over Shin Pond's thoroughfare. The departure time is 8:30 a.m., Saturday, October 11, 1980. Be sure to attach yourself to the correct trip; more than one trip will assemble here. Cars may be used on this trip but please be sure that each vehicle carries at least four people, has a full tank of gas, and has a useable spare tire fully inflated. Bring your own lunch. Note that pets, including dogs, are not permitted in Baxter State Park. The trip includes a 200-foot climb over a steep, forested scree slope and a 3-mile trailless walk down South Branch Ponds Brook. Stout walking shoes are essential; wet feet guaranteed for all but the most agile. Please stay with the leader and in as compact a group as possible.

Mileage

- 0.0 Site of Shin Pond House, facing northwest along Grand Lake Road. Trip B-4 also follows Grand Lake Road as far as the East Branch of the Penobscot River. Some of Neuman's road log for that trip as far as the Bowlin Pond Road is repeated here. You should refer to that road log (this guidebook) for more detail.
- 0.3 Roadside ledges are thin-bedded, crossbedded quartzite of the Grand Pitch Formation and porphyritic Rockabema Quartz Diorite.
- 0.4 Roadside ledges are medium- and dark-gray slate and quartzite of the Grand Pitch Formation.
- 0.5 Very light colored and fine-grained phase of the Rockabema Quartz Diorite and Grand Pitch slate.
- 1.5 T6R7 town line.
- 1.6 Road on right to Snowshoe and White Horse Lakes.
- 1.9 View of Sugarloaf Mountain straight ahead. The mountain is capped by a metadiabase sill. The fossiliferous Shin Brook Formation crops out on the slopes of the mountain beneath the sill.
- 2.6 Crommet Spring lunch ground.
- 3.1 Ledges to left of road are of the fossiliferous Shin Brook Formation.
- 3.6 Ledges on left are metadiabase sill that overlies the Shin Brook Formation.
- 5.7 Ledges on left are quartzite of the Grand Pitch Formation.
- 5.9 Bridge over Seboeis River.

- 6.7 Road on right to Scraggly Lake.
- 7.1 Beginning of long straight stretch of road with view down road of North Traveler and Bald Mountains, both held up by the Traveler Rhyolite.
- 10.4 Side road right to Hay Lake.
- 10.7 Side road left to Bowlin Pond.
- 11.1 T5R8 town line. Gradational contact between Seboomook Formation and Matagamon Sandstone crosses road near here.
- 11.8 Roadside ledges are of Matagamon Sandstone, as are all roadside ledges as far as the shore of Grand Lake Matagamon.
- 13.1 STOP 1. "Hurricane Deck." Overlook and exposures of Matagamon Sandstone. The Matagamon Sandstone here is in a northeast-trending structural basin, the Hay Mountain basin (Rankin, 1965). These exposures are very nearly on the axis of the basin, and the sandstone dips gently northeast. In good weather, one can obtain a fine view here of the mountains to the west and south. To the southwest, Mt. Katahdin is visible between Turner Mountain on the left and Traveler Mountain on the right. The long mass of Traveler Mountain is across the valley of the East Branch. Although The Traveler (the highest point of Traveler Mountain) is only 3,541 feet high, it rises 3,000 feet above the river. The bare conical peak of Bald Mountain is set against North Traveler Mountain. The last mountain to the right, barely visible from here, is Horse Mountain on the shore of Grand Lake Matagamon. C.H. Hitchcock (1861) referred to this as the mountain with the inelegant name. Turner and Katahdin are composed of Katahdin Quartz Monzonite, the others, of the Traveler Rhyolite.
- 14.7 Bridge over East Branch of the Penobscot River, a favorite for white-water canoeists. H.D. Thoreau (1950) extolled the joys of the East Branch after his 1857 trip down it. The store on the east side of the bridge was not there at the time of my last pre-1979 visit, that is, at the time of the 1966 NEIGC trip. Civilization creepeth into the Maine Woods. Road right on west side of bridge leads 0.5 mile upriver to Grand Lake Dam at foot of Grand Lake Matagamon. Good exposures of Matagamon Sandstone form east abutment of dam.
- 15.7 Baxter State Park Boundary. This is the largest State park in Maine and has an area of nearly 200,000 acres. 180-m cliffs of rhyolite forming Horse Mountain on left.
- 15.9 STOP 2. Park as close to the edge of the road as you can. Climb about 200 ft up steep slope to the base of cliffs. Be extremely careful in crossing scree slope. Remember others are behind you. The Pogy Member of the Traveler Rhyolite forms the

cliffs above. The Matagamon Sandstone underlies the scree slope over which we climb. The contact, defined as the sharp change in lithology from underlying obviously stratified rocks to rhyolite above, is more or less exposed at the top of the scree slope and dips 20° W. The top 6 m or so of the Matagamon contain scattered pebbles of felsite and beds of tuffaceous sandstone, indicating that some volcanic activity preceded the main body of rhyolite.

The basal 0.6 or 1.2 m of the massive rhyolite are composed of nonwelded tuff in which devitrified shards are clearly visible (fig. 1A). This nonwelded tuff grades up into welded ash-flow tuff that appears to make up most of the rhyolite of the Horse Mountain cliffs. Fragments of collapsed pumice are visible in the rhyolite about a meter above the base. Deformed and flattened shards are visible in a thin section (fig. 1B) collected from this locality 1.5 m above the base. Columnar joints may be seen in the cliff face. These are most obvious in the main part of the cliff to the south and are perhaps most easily seen from the road. Some are as much as 1.2 m in diameter and at least 12 m long.

If one traces the contact along the base of the cliffs, it is seen to be an irregular surface having relief of as much as 5 or 6 m. This irregularity may be due to scouring by ash flows, although in other places in the world, ash flows are known to have crossed unconsolidated material without disturbing that material.

- 16.3 STOP 3. Canoe landing on right opposite Maine Forest Service Camp. Known locally as Eastern Landing. Walk 0.1 mile ahead (north) along road. Roadcut in Pogy Member showing thin anastomosing dikes of sandstone from the underlying Matagamon Sandstone. Thicker clastic dikes have been found at the base of the cliffs on Horse Mountain and on the shore of Grand Lake Matagamon just ahead of us on the point. The largest clastic dike is about 6 m thick and at least 10 m long (as viewed from the bottom of the cliff). The clastic dikes provide evidence that the ash flows overrode unconsolidated sand of the Matagamon.
- 16.4 Gate house, Baxter State Park.
- 16.8 Road turns left away from lake and crosses ledges of rhyolite of the Pogy Member.
- 17.9 Cross unexposed, high-angle fault between Traveler Rhyolite and Seboomook Formation.
- 19.0 Trout Brook Farm, first cleared in 1837, is now a Baxter State Park campground. It produced hay for horses used in logging operations until the 1940's. C.H. Hitchcock stayed here in 1861. Rough side road, right, passes through farm

Figure 1. Photomicrographs of the Traveler Rhyolite. All are in plane-polarized light. A., B., and C. are of the Pogy Member and are in stratigraphic order from bottom to top. D., E., and F. are of the Black Cat Member and do not represent a stratigraphic succession.

- A. Nonwelded tuff, Pogy Member. Sample KT-153. Devitrified shards showing axiolitic texture and unruptured "bubbles" are visible. Stop 2 of field trip; altitude of about 900 feet at base of cliff on Horse Mountain about 0.5 m above the Matagamon Sandstone.
- B. Welded ash-flow tuff, Pogy Member. Sample KR-73. Phenocrysts are quartz and plagioclase. Flattened devitrified shards are visible; some are deformed around phenocrysts. Clast of siltstone is in upper right. Same locality as sample in figure 1-A about 1.5 m above the Matagamon Sandstone.
- C. Welded ash-flow tuff, Pogy Member. Sample KR-65. Texture suggests vague outlines of highly compacted devitrified shards. The dark mineral in the shards is probably celadonite. Altitude about 1,420 feet at top of cliff on Horse Mountain.
- D. Welded ash-flow tuff, Black Cat Member. Sample Kt-499. Phenocrysts of zoned plagioclase and altered augite. Highly compacted and devitrified pumice lumps impart the eutaxitic texture. The same texture on a larger scale is visible in the outcrop. From second unit above stream about 30 m downstream from Station 3 of traverse down South Branch Pond Brook.
- E. Welded ash-flow tuff, Black Cat Member. Sample KT-297. Highly compacted. Phenocrysts of plagioclase, augite, and opaque minerals. Growth aggregates of phenocrysts are visible. From ledges at an altitude of 1,160 feet on east bank of stream flowing north from Black Brook Mountains (Traveler Mountain quadrangle). This is about 2.8 miles west of South Branch Ponds campground.
- F. Welded ash-flow tuff, Black Cat Member. Sample KT-208. Phenocrysts of zoned plagioclase, augite, biotite, and opaque minerals. Fayalite is also present in the thin section. Extreme compaction of pumice lumps. Rotated phenocrysts indicate that the material flowed after compaction and welding so that texture resembles that of a lava. From outcrop at an altitude of about 2,240 feet along an unnamed stream that flows east from The Traveler toward Haskill Deadwater of the East Branch.

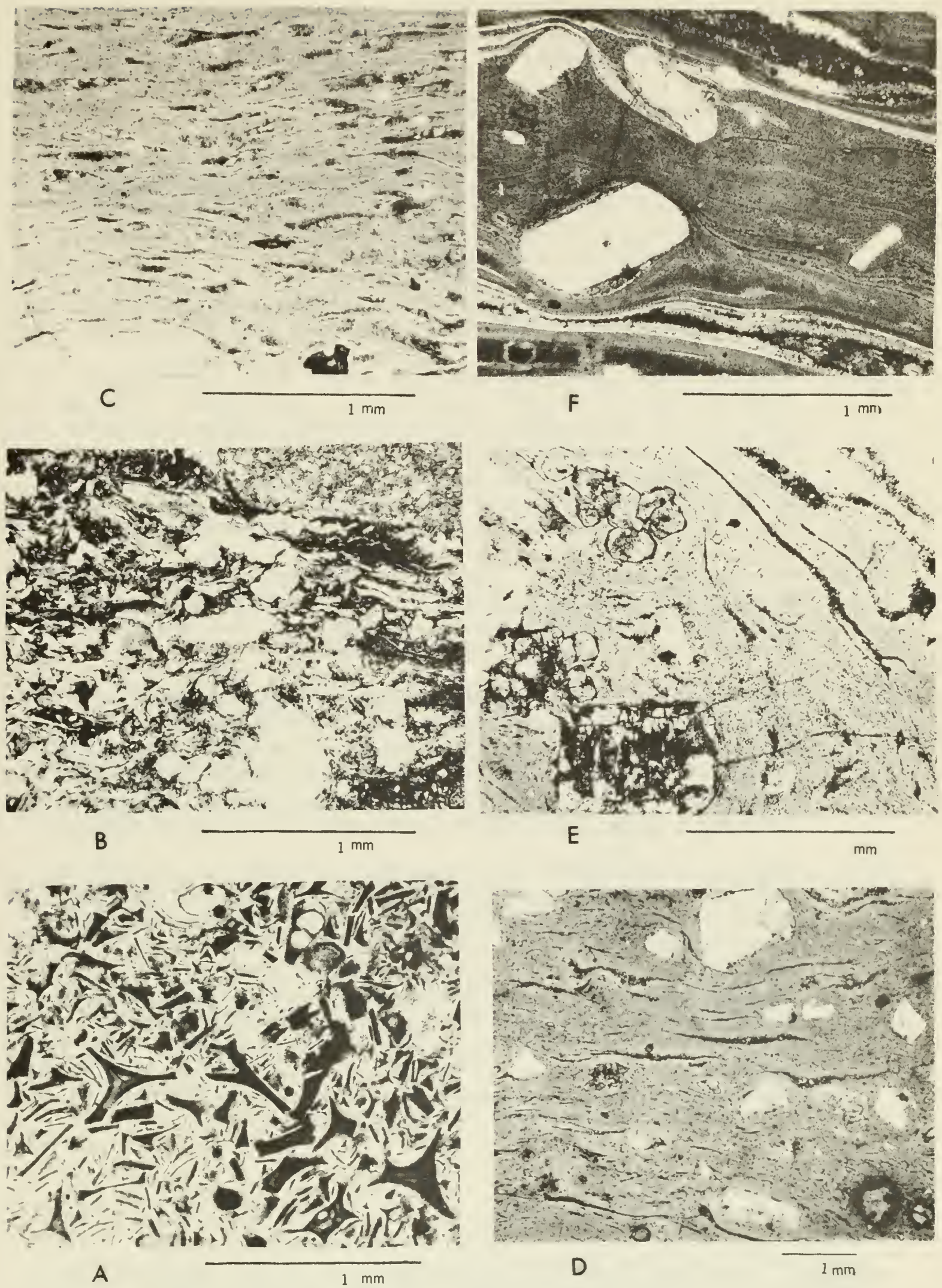


Figure 1.

and continues to Webster Brook at the head of Grand Lake Matagamon, crossing enroute some well-exposed open folds in the Seboomook Formation.

- 19.5 Trout Brook on right parallel to road.
- 20.0 Sharp turn left. Ledges of Seboomook Formation in woods to left. Excellent exposures of the Seboomook Formation just upstream from the adjacent right-angle turn of Trout Brook. Graded bedding, refracted cleavage, and many small folds have been observed in these exposures.
- 20.2 Parking area left for trail to the delightful lakes of the Deadwater Mountains.
- 20.5 STOP 4. Park along main road and walk 0.1 mile along side road to site of old K.P. wooden dam on Trout Brook. The dam is built on ledges of brecciated Black Cat Member, Traveler Rhyolite. The high-angle fault bounding the rhyolite on the north crosses the stream just below the dam. A thin wedge (9 to 12 m) of much fractured Matagamon Sandstone is north of the fault. Beyond this is the Seboomook Formation.
- 22.9 Crossing of Dry Brook. Spectacular columnar jointing in the upper distinctive part of the Black Cat Member about a mile upstream. The columns are deformed into an open z-fold by a normal fault of about 0.5-m displacement.
- 23.4 The Crossing. STOP. Leave appropriate number of cars so that drivers may later be ferried to South Branch Ponds to retrieve remaining cars. Considerable doubling up will be necessary, but it is a short drive and no one wants to walk both ways. Take lunches with you. After leaving some cars, proceed up side road left to South Branch Ponds Campground.
- 25.6 STOP 5. South Branch Ponds Campground. Park cars in parking area at entrance to campground and walk to shore of pond for lunch. After lunch we will leave the cars here and walk down South Branch Ponds Brook to The Crossing. There is no trail, the distance is nearly 3 miles, and it is practically impossible to make the trip with dry feet. Please stay with the group. We must leave the campground by 1:30 p.m., and we must all be at The Crossing no later than 4:00 p.m. The sketch map (fig. 2) is traced from an aerial photograph, so the scale is approximate. The log of the walk is by numbered stations on the map (fig. 2), not distance.

South Branch Ponds Brook Wade.

- 1. Shore of Lower South Branch Pond.
South Branch Ponds occupy a glacial valley breaching a large anticline in ash-flow sheets of the Black Cat Member of the Traveler Rhyolite. On Black Cat Mountain to the west (right, looking up the

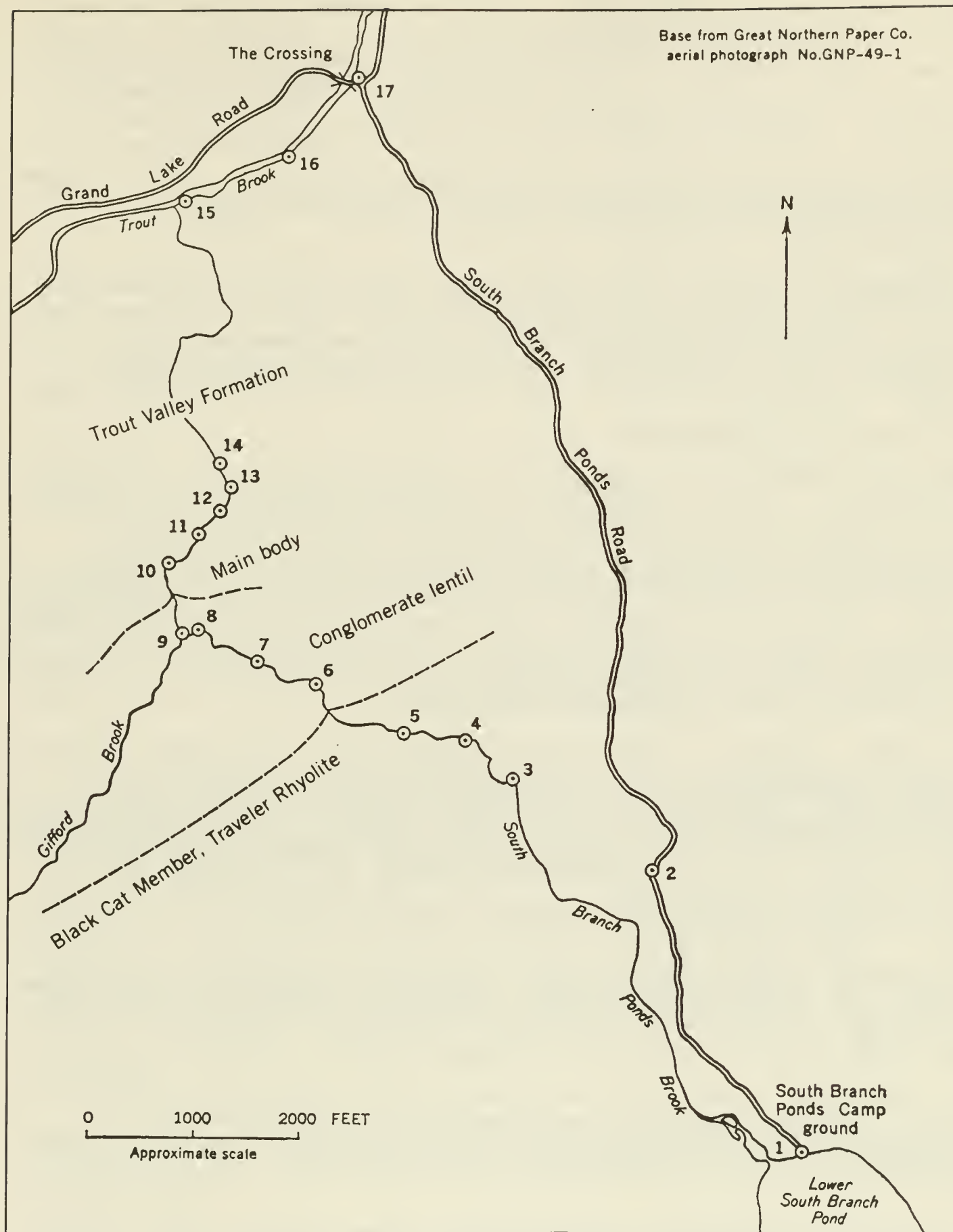


Figure 2. Sketch map of the South Branch of Ponds Brook area. Numbers refer to localities described in the text.

lake away from the campground), ash flows strike northeast and dip moderately northwest. On Traveler Mountain to the east (left), ash flows strike northwest and dip moderately northeast. The attitude of these flows controls the northwest pattern of ridges on Traveler Mountain. Neither the summit of The Traveler nor North Traveler is visible from the lake shore. Mt. Katahdin is visible over the end of the Upper Pond from the ridge north of the campground. Retrace route out of campground past parking area and along road toward The Crossing.

2. Reassemble on road at top of long hill (about 0.6 mile from shore of lake). Trail right, which we will not take, leads to the Ledges, open ledges of the Black Cat Member overlooking South Branch Ponds. Columnar joints are distinct. Turn left downslope through open woods. South Branch Ponds Brook is reached in about 0.2 mile. Turn right and walk downstream.
3. Exposures of Black Cat Member of the Traveler Rhyolite. Note pattern of concentric joints in outcrop on corner at stream level. Actually there is more than one center about which joints are concentric, giving rise to a pattern of intersecting curving joints. Distinct columnar joints of small diameter are present above the stream on east bank. Above this and slightly downstream is another flow unit having columnar joints of larger diameter. Note the compaction foliation in the unit (eutaxitic texture) brought out by the presence of very thin lenticular bodies (collapsed pumice lumps).
4. First of a series of joint surfaces across which the stream flows. Close examination will show that these are dip surfaces of ash flows. Compaction foliation is roughly parallel to the surfaces, and, in places, rather crude columnar joints are roughly perpendicular to the surfaces. The ash flows strike about N. 70° E. and dip 30° N. Local areas of crosscutting breccia are also present in this outcrop on the left bank of stream.
5. Last of the series of dip slopes. Excellent swimming holes at bottom of falls. Compaction foliation is visible on a number of steep joint surfaces.
6. In stream bed on east bank just upstream from steep gravel bank at corner. Lowest exposure of conglomerate of the Trout Valley Formation. You, too, might call this a Pleistocene till upon first encounter.
7. Jointing cuts cobbles in conglomerate. First clue that this is not a Pleistocene till. Some cobbles are offset along the joints. Note that clasts (pebbles, cobbles, and boulders) are well rounded and that all of them are rhyolite. The clasts are so weathered that many of them can be broken apart by hand. The weathering may date from the Devonian Period. The conglomerate is crossbedded and dips gently north, away from the volcanic rocks.

8. About 10 m of conglomerate exposed in the canyon wall. Where is the contact with the overlying till? Note sandstone bed in the conglomerate near top of exposure and lenses of black sandy carbonaceous shale near bottom.
9. Junction with Gifford Brook.
10. Large exposure at curve of stream on left bank. Coarse conglomerate no longer dominant. We are now above the basal conglomerate lentil and in the main body of the Trout Valley Formation. Numerous black chert lenses are visible and some have a vague internal structure. Professor E.S. Barghoorn, of Harvard University, has identified one of these as Prototaxites, which is generally regarded as of algal affinities. Also present are siderite concretions and thin beds of sideritic ironstone.
11. Upstream: sill of intermediate rock and a 15-cm-thick bed of ironstone. Downstream: lens of carbonaceous black shale, from which I first collected plant fossils in July 1955.
12. Light-gray-green fine-grained intermediate dike about 1 m thick. Trends N. 30° W. and dips 40° N. Note chilled contact against the sedimentary rocks.
13. This is "locality 4" of Dorf and Rankin (1962), from which the best specimens of flattened spiny stems of Psilophyton were collected.
14. Gently dipping sill of intermediate rock about 3 m thick. Plant remains can be found in nearly every outcrop of sedimentary rocks. Reassemble here for half-mile walk out to Trout Brook. The group must stay together from this point on. We will not follow the stream, but will cut across country toward The Crossing.
16. "Locality 1" of Dorf and Rankin (1962). Long outcrop of gently dipping, interbedded sandstone and shale of the Trout Valley Formation. Sandstone is calcareous and current bedded. Rather well preserved plant fossils have been recovered from some of the fine-grained sandstone. Eurypterid scales also were found at this outcrop. A fault of unknown magnitude cuts the southwest end of the exposure.
17. The Crossing. Poorly preserved but large plant fossils occur in the bridge abutment. Drivers will be ferried to South Branch Ponds to recover cars.

End of trip. Return to Shin Pond and proceed to Presque Isle for evening program.

TRIP B-4

ITINERARY

THE CORE OF THE WEEKSBORO-LUNKSOOS LAKE ANTICLINE, AND THE ORDOVICIAN,
SILURIAN, AND DEVONIAN ROCKS ON ITS NORTHWEST FLANK

Robert B. Neuman
U.S. Geological Survey, Washington, DC 20560

Topographic quadrangle maps:

15-minute

Shin Pond

Traveler Mountain

2-degree

Presque Isle

Assemble at site of Shin Pond House, now burned, Shin Pond, Maine ready for departure at 8:30 A.M., Saturday, October 11. Road limitations dictate that heavy-duty vehicles be used. Please insure that (1) each vehicle carries at least 4 people, (2) at the start the fuel tank of each car is full, and (3) the spare tire is inflated and useable. Most stops will be off the road, examining sections in the woods along streams. It will be in the best interests of all participants to keep the group as compact as possible. When the leader decides adequate time has elapsed for examination of the exposures, he will return to the cars and proceed to the next stop. Therefore, keep alert to the movements of the group, and do not risk being left in the woods.

Mileage

- 0.0 Site of Shin Pond House, facing northwest.
- 0.3 Roadside ledges are thin-bedded, crossbedded quartzite of the Grand Pitch Formation and fine-grained Rockabema Quartz Diorite.
- 0.4 Roadside ledges are medium- and dark-gray slate and quartzite of the Grand Pitch Formation.
- 0.5 Very light colored and fine-grained phase of the Rockabema Quartz Diorite and Grand Pitch slate.
- 0.9 T5R7 town line.
- 1.3 T6R6 town line.
- 1.5 T6R7 town line.
- 1.6 Road on right to Snowshoe and Whitehorse Lakes.
- 1.9 View of Sugarloaf Mountain straight ahead. The mountain is capped by a metadiabase sill in a syncline plunging northeast. Beneath the sill, on the slopes of the mountain is the Shin Brook Formation. The best fossils from the Shin Brook were found on the easily climbed southern slope at 1,500 feet altitude.

2.6 Spring and lunchgrounds.

- 2.7 **STOP 1. SHIN BROOK FORMATION (fig. 1):** Stop cars near Crommet Spring lunch ground; walk down road to left (west), (recently relocated from alignment shown on sketch map) to bridge over Shin Brook. Roadside exposures and east bank of brook are Grand Pitch Formation: medium- to dark-gray slate and thin-bedded, laminated, fine-grained light-gray quartzite, some crossbedded. Note tight, steeply plunging folds, especially in streambank.

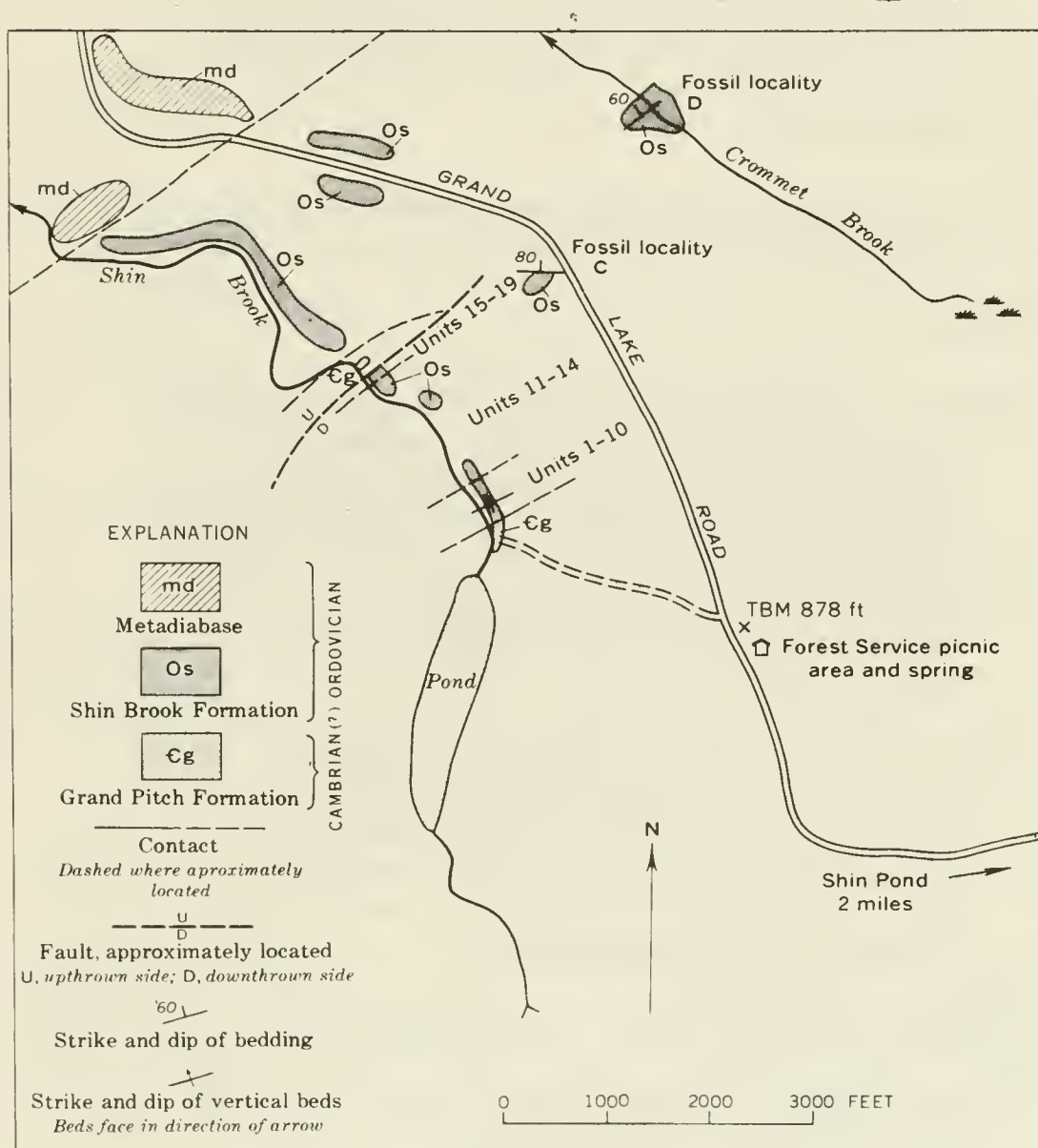


Figure 1. Geologic sketch map of the vicinity of the type section of the Shin Brook Formation; units 1-19 described in text. Outlined areas indicate outcrops. From U.S. Geological Survey Bulletin 1181-E, fig. 2.

The base of the Shin Brook Formation is a spotted slate which is about 60 cm (2 ft) thick. The type section measured here (taken from U.S. Geol. Survey Bull. 1181-E, p. E-6) consists of:

	Thickness	
	(Meters)	(Feet)
Shin Brook Formation: 280 m (902 ft) measured		
19. Tuffaceous sandstone and siltstone in graded layers 8-30 cm (3-12 in.) thick, with coarse-grained sandstone in the basal part and finely laminated siltstone at the top; siltstone more abundant than sandstone; unit includes two layers of crystal tuff 16 cm (6 in.) and 60 cm (2 ft) thick, respectively.....	10	35
18. Crystal tuff, greenish-gray; crystals are green altered plagioclase.....	3	10
17. Covered.....	1	3
16. Tuffaceous sandstone, grit, and conglomerate; finer grained part is well laminated, coarser part not laminated and includes fragments of porphyritic and nonporphyritic fine-grained igneous rock; ledge in stream-bed has distorted bedding structures that suggest deformation prior to lithification; base concealed.....	2	7
15. Tuffaceous sandstone, fine- to medium-grained, calcareous; strongly sheared with weathered pits that may have been concentrations of fragmental fossils; fossils largely brachiopods, too strongly deformed to be identified.....	5	18
14. Covered.....	20	70
13. Crystal tuff, greenish-gray; with scattered angular cognate rock fragments 1-15 cm ($1\frac{1}{2}$ -6 in.) in average diameter; crystals of both matrix and fragments are green altered plagioclase; no primary layering seen; quartz veins abundant.....	6	20
12. Covered.....	115	375
11. Crystal tuff; light-green altered plagioclase phenocrysts in a darker aphanitic matrix; fractured.....	15	45

10. Covered.....	30	95
9. Tuff, fine-grained, light-greenish-gray; abundant carbonate; strongly sheared, with no bedding structures preserved.....	10	30
8. Covered.....	10	30
7. Volcanic conglomerate, with granules of aphanitic volcanic rock and dark slate, strongly sheared; bedding obliterated.....	15	50
6. Tuffaceous sandstone, gray, medium- and fine- grained; abundant carbonate; bedding obliterated by strong shearing.....	6	20
5. Covered.....	10	30
4. Volcanic granule conglomerate and coarse- grained sandstone, light-gray, strongly sheared.....	8	25
3. Sandstone and conglomerate, interbedded, with conglomerate beds 25-50 cm (10-20 in.) thick sandstone somewhat thinner; angular to subrounded fragments as much as 4 cm (1 1/2 in.) in average diameter include volcanic rocks and fine-grained quartzite....	12	35
2. Phyllite, light-gray, probably tuffaceous.....	1	2
1. Slate, dark-gray, with small (1/4 - 1/2 mm) white grains (altered plagioclase?) with rhombic outline abundant.....	1	2

Return to vehicles and proceed northwest on Grand Lake Road.

- 3.1 Ledges to left of road include fossil locality C of fig. 1 from which large specimens of Orthambonites robustus Neuman, deformed Platystrophia sp., and bryozoans were collected. Fossil locality D, shell beds of O. robustus, is about 500 meters (1,500 ft) to the northeast in Crommet Brook.
- 3.6 Ledges on left are metadiabase of the sill that overlies the Shin Brook Formation.
- 4.9 STOP 2. GRAND PITCH FORMATION AT SHIN FALLS: Walk south along old road about 500 meters (2,500 ft) to second small road to left (rushing water is plainly audible at this point). Turn left (to east) and follow this road about 1,200 feet to Shin Brook at bridge (do not cross bridge); then follow Shin Brook westward, downstream. The first large ledges are greenish-gray, fine-grained quartzite with interbedded gray and red slate. Crossbedding and graded

bedding are not as conspicuous as they are in the next exposures downstream. Exposures at the upper cascades of the falls have somewhat more slate and thinner quartzite layers than here. Graded bedding and crossbedding indicate that beds face in the same direction for only short distances, and then are abruptly reversed. Please do not descend waterfalls, but return to road and cars from above log sluice.

- 5.7 Ledges on left are quartzite of the Grand Pitch Formation.
- 5.9 Bridge over Seboeis River.
- 6.7 Scraggly Lake Road, TURN RIGHT.

STOP 3. SEBOOMOOK FORMATION AT SAWTELLE FALLS: Ledge on the north bank of Sawtelle Brook is fine-grained sandstone identical to that of the Matagamom Sandstone, one of several that occur throughout the Seboomook. Walk eastward along old road and trail about 2,000 feet to falls. The falls afford an especially informative exposure of the Seboomook Formation as large areas of bedding surface are visible. The ripplemarked fine-grained sandstone is especially interesting, as it reveals in plan view what is seen as small-scale crossbedding at the base of graded sets in cross section. Note the regular orientation of the ripplemarks and their elongation parallel to the intersection of bedding and cleavage. The plunge pool and its downstream extension follow the trough of a syncline, the beds on the opposite side of the stream dipping steeply northwest. This is the only fold seen in the area that has a horizontal axis.

Reverse direction; return to Grand Lake Road.

- 9.7 Grand Lake Road. TURN RIGHT.
- 10.6 Roadside ledge of Seboomook Formation.
- 11.0 Roadside ledge of Seboomook Formation.
- 11.3 Forest Service Camp, road right to Hay Lake; glimpse of lake from the highway.
- 11.5 Bowlin Pond Road. TURN LEFT.
- 12.8 Weathered exposure of Seboomook Formation.
- 13.3 Roadbed "pavement" of Seboomook Formation.
- 13.4 Woods road left. Before parking for Stop 4 prepare to reverse direction of caravan by proceeding past this road far enough so that last vehicle can back into it, leaving sufficient space so that all vehicles can be accommodated.

12.7

STOP 4. SILURIAN SEQUENCE AND CONTACT WITH GRAND PITCH FORMATION. Follow woods road on which vehicles are parked about 1,500 feet east and south to crossing of Bowlin Brook. In roadbed at and near crossing note red shale (Grand Pitch) interspersed with pebble conglomerate (Silurian), probably tectonically. Proceed about 300 feet south down small gully to its mouth at Bowlin Brook. Red and green slate and siltstone of the Grand Pitch form the southeast wall of the gully; southward along the brook for several hundred feet are nearly continuous exposures of red, green, and gray slate and greenish-gray, fine-grained, laminated, thin bedded quartzite typical of parts of the Grand Pitch.

Debris on the gully floor, about 10 feet wide, covers the trace of the contact between the Grand Pitch and the overlying Silurian rocks, which are gray and dark-gray slate and slaty siltstone containing thin seams of coarse-grained sandstone exposed in waterfalls. Note the contrast of deformation on opposite sides of the covered interval; also note that the lineation formed by the intersection of cleavage with bedding surfaces of the Silurian rocks plunges to the northeast. (Although the contact between the Grand Pitch and Silurian rocks here has previously been considered to be an unconformity, perhaps slightly faulted due to contrasts of structural competency of the rocks juxtaposed here, geologic relations in this vicinity, and the variety of units northwest of the Grand Pitch along this strike belt, some almost certainly overthrust by the Grand Pitch, suggest the possibility of major fault movement at this level, perhaps a decollement.)

About 50 feet west of the brook are beds of crudely graded quartzose pebble conglomerate and coarse-grained sandstone that are probably covered beneath the waterfalls in the brook. Above the waterfalls are about 30 feet of red slate and siltstone which are in turn overlain by about 50 feet of interbedded conglomerate, sandstone, and siltstone; these conglomerates contain Pentamerus and other Silurian brachiopods.

After examining this section, return to the Bowlin Pond Road for examination of the remainder of the section.

The following section (revised from U.S. Geol. Survey Prof. Paper 524-I) was measured in the woods west of the road (fig. 2).

	Thickness (Meters) (Feet)
Seboomook Formation, 51 m (170 ft) measured, of which 28 m (95 ft) is exposed):	

12. Siltstone, dark-gray, in beds 10-15 cm
(4-6 in.) thick; basal parts formed

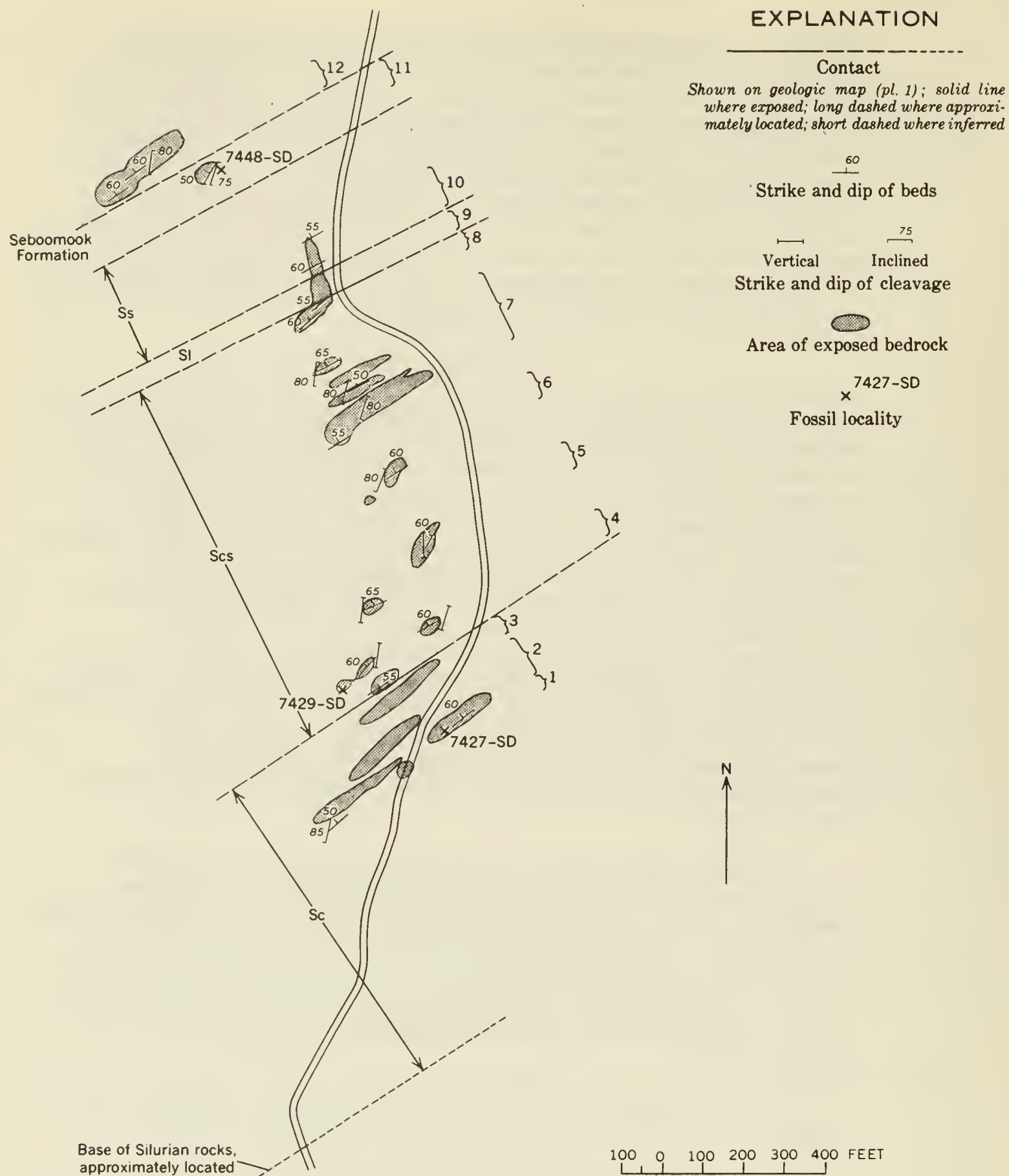


Figure 2. Location of exposures and units of Silurian sequence along Bowlin Pond Road, Shin Pond quadrangle, described in text. Sc-conglomerate, Scs-lower calcareous siltstone, Sl-limestone, Ss-upper calcareous siltstone. From U. S. Geological Survey Professional Paper 524-I, figure 2.

of fine-grained crossbedded sandstone 1-3 cm ($1\frac{1}{2}$ - $1\frac{1}{2}$ in.) thick.....	20	70
Covered interval.....	8	25
11. Siltstone, gray, sandy, noncalcareous; bedding obscure but shown in subtle contrast between more and less sandy parts; scattered small brachiopods, <u>Metaplasia?</u> sp. and <u>Plectodonta</u> sp.....	8	25
Covered interval.....	15	50
Upper Silurian Rocks, at least in part, 273 m (820 ft) measured, of which 153 m (515 ft) is exposed):		
Covered interval.....	55	180
10. Siltstone, gray, calcareous; bedding surfaces even, planar, at 5-25 cm (2-10 in.) intervals; widely scattered comminuted organic debris.....	25	85
9. Calcarenite, largely pelmatozoan debris; silty partings irregular and discontinuous; stromatoporoids and favositid corals as much as 15 cm (6 in.) in cross section.....	15	50
8. Siltstone, calcareous; contains silty lime- stone nodules a few centimeters (inches) in diameter, irregularly distributed.....	8	25
Covered interval.....	20	70
7. Siltstone, calcareous, gray, largely massive, but with a few scattered silty limestone nodules 3-5 cm (1-2 in.) in diameter, in beds 12-25 cm (5-10 in.) thick, separated by fine-grained lime- stone beds 2-5 cm (1-2 in.) thick, some of which are cut into boudins by partitions of siltstone parallel to slaty cleavage.....	50	170
Covered interval.....	20	70
6. Siltstone, gray, calcareous, in faintly laminated beds separated by prominent partings at intervals of 5-25 cm (2-10 in.); brachiopods fragmentary, scarce, <u>Monograptus</u> sp. rare.....	20	65

Covered interval.....	25	85
5. Siltstone, calcareous, like unit 6 above.....	15	50
4. Siltstone and fine-grained sandstone, calcareous, light-gray, mostly in well- defined graded beds 12-24 cm (8-10 in.) thick, some prominently laminated. A dark-gray noncalcareous siltstone bed 1 m (3 ft) thick, 20 m (60 ft) above the base of the unit contains 17 genera of brachiopods including <u>Cyrtia</u> sp., <u>Eospirifer</u> cf. <u>E. radiatus</u> (Sowerby), <u>Howellella</u> sp., <u>Merista</u> sp., and <u>Sphaerirhynchia</u> sp., trilobites and corals.....	20	70
Lower Silurian Rocks, 54 m (175 ft) measured, of which 36 m (115 ft) is exposed)		
3. Pebble and granule conglomerate, sandstone, and siltstone, calcareous, in well- defined graded beds 12-25 cm (4-8 in.) thick; fragments of brachiopods and corals.....	14	45
Covered interval.....	10	35
2. Conglomerate with rounded pebbles and granules, mostly of quartz, but with abundant felsitic volcanic rocks and tabular fragments of slate and siltstone as much as 15 cm (6 in.) long; beds poorly defined, about 1.5 m (4 ft) thick; scattered brachiopods including <u>Stricklandia lens ultima</u> Williams, tabulate and rugose corals.....	20	65
Covered interval.....	8	25
1. (Beds exposed in floor of Bowlin Pond Road) Siltstone, greenish-gray, and fine- grained sandstone; 2 beds of granule conglomerate, each about 25 cm (10 in.) thick.....	2	5
Beds below not exposed; distance to Grand Pitch Formation.....	175	575

Return to Grand Lake Road.

17.1 Grand Lake Road. TURN LEFT.

18.2 Enter Traveler Mountain Quadrangle.

- 19.3 STOP 5. "Hurricane Deck." Overlook and exposures of Matagamon Sandstone. For description and discussion see Stop 1 of trip B3 (Rankin, this guidebook).

Continue westward on Grand Lake Road.

- 20.8 Bridge over East Branch of the Penobscot River. Immediately after crossing bridge, TURN LEFT. Road skirts campground; keep right, away from campground.
- 21.0 Culverts (if luck is with us) over backwater. Woods road follows approximately old Eagle Lake Tote Road.
- 26.0 STOP 6. Haskell Rock; park vehicles as space permits here and half mile to south.

The purpose of this stop is to examine the Upper Ordovician succession along the west side of the river between Pond Pitch and Haskell Rock. Although fossiliferous Upper Ordovician rocks (equivalent to the British Ashgill Series) are known from a few other places in Maine (Lobster Mountain Volcanics of Boucot and Heath (1969); Blind Brook Formation of Hall (1970); and at several places in the vicinity of Ashland), this section has so far proved unique for its completeness, the apparent amount of time that it represents, the abundance and diversity of its fossils, the variety of rocks that it includes, and its relatively simple structure. The presence of this thick section (1,200 m, 3,500 ft) here, compared to the total absence of equivalent rocks in the Bowlin Brook exposures of Stop 4, three miles to the northeast in the same outcrop belt, may be due as much to structural causes as to stratigraphic ones. One possibility not hitherto entertained is that faulting at this level rearranged bodies of rock so their present distribution does not reflect their original relations.

Figure 3 represents exposures on both sides of the river constructed from a 1:2,400-scale tape and compass survey made in 1976. The exposures on the west side that should be seen during this trip are listed below, from south to north.

1. Basalts at Pond Pitch, identified by Rankin (1961) as olivine basalt with rare pillows, about 100 m (300 ft) thick; quartzite of the Grand Pitch Formations crop out about 75 m (200 ft) to the south of the southernmost basalt outcrop on the east side of the river, but the contact between them is not exposed. Small pillows near the top of the basalt, immediately below the contact with the overlying sediments, are visible when the river level is low.
2. Lower siltstone, about 25 m (80 ft) exposed; dark-gray siltstone, bedding obscure in most places, strongly cleaved; fossils obtained from small roadside ledge on west side of

PLANT FOSSILS (PSILOPHYTES) FROM THE DEVONIAN
TROUT VALLEY FORMATION OF BAXTER STATE PARK

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INTRODUCTION

Although small in geographical extent, the Trout Valley Formation contains one of the richest early-land-plant deposits in the world. Dorf and Rankin first described the flora and defined the formation in 1962. Since then, Andrews (Biological Sciences, Univ. Connecticut) and his students in a series of papers over the years have reported numerous taxa from these strata. A recent paper with reconstructions and illustrations of both micro- and mega-plant fossils has summarized their work up to now (Andrews et al., 1977).

The Trout Valley Formation consists of relatively undeformed continental strata lying in the trough of a synclinorium whose axis strikes east-northeast in the Traveler Mountain Quadrangle, Maine. The formation contains the youngest sedimentary rocks in the area. Their maximum exposed thickness is about 1500 feet and they outcrop over an area 1.5 by 8 miles in the valley of Trout Brook northwest of Traveler Mountain in Baxter State Park (Dorf & Rankin, 1962). These clastic rocks are a heterogeneous assemblage of shale, siltstone, sandstone and conglomerate. The formation unconformably overlies the Traveler Rhyolite which, in turn, rests conformably on the Matagamon Sandstone (Dorf & Rankin, 1962). The marine fauna of the Matagamon Sandstone is of Becraft-Oriskany age (=Siegenian of Europe, Lower Devonian; Rankin, 1965).

Dorf and Rankin (1962) date the Trout Valley Formation as Early Devonian based on the flora. Because of lithological similarities, Rankin (1961) had previously correlated the formation with the marine beds of the Tomhegan Formation to the southwest. Boucot et al. (1964, p. 94) date the Tomhegan Formation as Schoharie (=Emsian, Lower Devonian). A palynological analysis by Andrews et al. (1972) suggests an Emsian/Eifelian age (late Lower/early Middle Devonian). An absolute date for the underlying Traveler Rhyolite is given as 360-10 m.y. (Bottino et al., 1966) which suggests a much younger age, Givetian/Frasnian (late Middle/early Upper Devonian). In their summary paper based on both plant megafossils and paleopalynology, Andrews et al. (1977) suggest that the formation is Emsian (late Lower Devonian). Finally, the recent discovery of the hitherto Middle Devonian lycopod, Leclercqia complexa, in the formation adds to the dating problem and favors a Middle Devonian age (Kasper & Forbes, 1979).

The majority of species described in the Trout Valley Formation belong to a group of extinct plants which previously had been called 'psilophytes'. In 1968 Banks re-classified the psilophytes into three distinct Subdivisions, one of which, the Trimerophytina, is abundantly represented in the Trout Valley flora. In addition to this Subdivision, lycopods--well known from the 'scale trees' of the Carboniferous--are also present in the formation, however, only as small herbaceous forms.

PSILOPHYTE CLASSIFICATION

Early land plants have been called 'psilophytes' after the genus, Psilophyton, described by J. W. Dawson in 1859 from the Gaspé Peninsula, Quebec. Dawson's historic Psilophyton princeps was one of the first-reported early land plants from North America. The name Psilophyton, i.e., naked-plant, correctly characterizes the group as plants without leaves. For many years after, paleobotanists assigned the numerous and varied early Devonian fossil plants to this single category, the psilophytes. It was becoming readily apparent, however, that the psilophytes were a heterogeneous assemblage and an artificial taxon. In 1968 and more recently, 1975, Banks re-classified the psilophytes into three major groups based on structural and reproductive features.

Subdivision Rhyniophytina:

The Subdivision name comes from the genus Rhynia described from the silicified Devonian peat beds near Rhynie, Scotland. The Subdivision Rhyniophytina contains small leaf-less plants a few decimeters in height with dichotomously forking stems and single sporangia borne at the branch tips. The sporangia or spore sacs are generalized as being spindle-shaped and splitting lengthwise. A cross-sectional view of 'rhyniophyte' stem anatomy shows a small centrally-located strand of conducting tissue round in outline with the first-formed cells in the center (centrarch).

The rhyniophytes are the most primitive vascular land plants known--primitive from both a botanical and geological standpoint. Morphologically and anatomically they are the smallest and simplest of plants. Geologically they contain the oldest unquestionable vascular plant genus, Cooksonia. Cooksonia was described by the paleobotanist, Lang, in 1937 from latest Silurian (=Downtonian) strata of Wales. Banks (1974, 1975) has discussed the controversial topic of the 'oldest land plant' in two recent papers.

Subdivision Zosterophyllophytina:

The Subdivision Zosterophyllophytina is the second group to which a portion of the psilophytes have been assigned. The name comes from the early described genus, Zosterophyllum (Penhallow, 1892) from the Lower Devonian of Scotland. The 'zosterophylls' are larger plants than the rhyniophytes and may have attained heights up to a meter. They also had dichotomously forking stems, however, 'overtopping' was common. Overtopping occurs when one limb of a dichotomy continues as the main shoot and the other limb is restricted in growth as a side branch. Even though all parts of the plant are equivalent botanically speaking, the plant appears to have a 'main stem' and 'side branches'--the pseudomonopodial habit. Overtopping and an increase in size of the conducting tissue permitted plants to grow taller. The conducting tissue in cross-sectional view is much larger, elliptical in outline and has the first-formed cells at its periphery (exarch).

The key to identification of the zosterophylls is the location of their spore sacs, sporangia. The sporangia are attached along the stem--in a lateral position--rather than at the tips or terminal position as in the rhyniophytes. The spore sacs are kidney-shaped (reniform) and open by a distal suture. Finally, the zosterophylls may either be unornamented or their stems may display a variety of surface ornamentation or enations. Enations vary from species to species and are important in identification. They range in form from multicellular spines or glands to deltoid tooth-like emergences.

Subdivision Trimerophytina:

The third group to which a portion of the psilophytes have been assigned is the Subdivision Trimerophytina (Banks, 1968, 1975). The majority of the Trout Valley specimens belong to this category. This Subdivision contains the largest plants of the three. The 'trimerophytes' were probably over a meter tall, with a robust stem in which overtopping was so pronounced that specimens display a distinct main stem and side branches. These plants had much more conducting tissue than members of the previous two groups. This, along with thick-walled outer cortical cells, supported these taller plants. The conducting tissue was circular in cross section and the first-matured cells were in the center (centrarch). The location, shape and dehiscence of the sporangia are diagnostic. The spindle-shaped or fusiform sporangia are located at the ends of the side branches and form dense clusters rather than being solitary as in the rhyniophytes or scattered as in the zosterophylls. The spore sacs opened by means of a longitudinal slit. Some of the trimerophytes were unornamented, while others bore hair-like, gland-like or spine-like processes on their stems--important characteristics in field

identification.

FLORA OF THE TROUT VALLEY FORMATION

Plant fossils from the Trout Valley Formation were first described and illustrated by Dorf and Rankin in 1962. These significant but fragmentary specimens were assigned to six taxa of early land plants. This initial report was important in making geologists aware of the presence of plant fossils in northern Maine and their potential use in dating and correlating the isolated continental deposits in the region.

In 1964 and 1965 Forbes (Univ. Maine, Presque Isle), Mencher (City College, C.U.N.Y.) and Schopf (U.S.G.S., Columbus) guided Andrews to several plant localities in northern Maine and encouraged him to study the plant fossils of the area. In 1968 Andrews et al. reported a new species of Psilophyton, P. forbesii, named after its discoverer.

In 1969 Gensel et al. described a new plant, Kaulangiophyton akantha, which appears to be intermediate between the zosterophylls and the lycopods. This is an important addition to the accumulating evidence that the lycopods arose from the zosterophyll line.

Six years after the initial introduction to the area, Andrews and Kasper (1970) published a summary report on the flora. The text was accompanied with illustrations of specimens and reconstructions of plants. The age of the formation, based on plant fossils, was given as middle or upper Lower Devonian.

In 1972 a new trimerophyte, Pertica quadrifaria, was described by Kasper and Andrews based on exceptionally complete compression/impression specimens. The material permitted an accurate reconstruction of the plant providing information on its size, arrangement of side branches on the stem and the branching patterns in both fertile--sporangium bearing--and sterile branches.

In 1974 Kasper et al. described two new species of Psilophyton, P. dapsile and P. microspinosum, besides presenting additional information on the previously reported species, P. forbesii and P. princeps. This paper illustrated for the first time trimerophyte remains from the Fish River Lake Formation in the Eagle Lake/Saint Froid Lake area (Winter-ville and Eagle Lake Quadrangles) of northern Maine.

A detailed review of the megaplant fossils along with the first illustrated analysis of the microflora was presented by Andrews et al. in 1977. This is the best summary article, to

date, containing numerous photographs and reconstructions. Again, an age of either "...late Early Devonian or earliest Middle Devonian...." is suggested (Andrews et al., 1977, p. 283).

Finally, Kasper and Forbes (1979) presented the first detailed report of a lycopod from the formation. Leclercqia complexa, although preserved in a fragmentary condition, is readily identified because of the unique morphology of its leaves. Up until now this lycopod has been known only from the Middle Devonian, so it adds to the controversy regarding the age of the Trout Valley Formation.

A revised list of plant mega- and microfossils is presented below. Each taxon is followed by one or two selected references recording its presence in the Trout Valley Formation and, if illustrated, by the plate and figure numbers. For comments on Dorf and Rankin's (1962) original determinations see Andrews et al. (1977, p. 272).

Megafossils:

Subdivision Rhyniophytina

Taeniocrada sp. -- Dorf & Rankin, 1962, Pl. 140, Fig. 9; Andrews et al., 1977, Pl. VI, Fig. 4.

Subdivision Zosterophyllophytina

Sawdonia ornata -- Dorf & Rankin, 1962, Pl. 140, Fig. 1-4.

Kaulangiophyton akantha -- Gensel et al., 1969.

Subdivision Trimerophytina

Psilophyton forbesii -- Andrews et al., 1968; Kasper et al., 1974, Fig. 21-26.

P. dapsile -- Kasper et al., 1974, Fig. 5-9.

P. microspinosum -- Kasper et al., 1974, Fig. 13-19.

P. princeps -- Kasper et al., 1974, Fig. 28-33.

P. sp. -- Andrews et al., 1977, Pl. II.

Pertica quadrifaria -- Kasper & Andrews, 1972.

Subdivision Lycophytina

Drepanophycus sp. -- Andrews & Kasper, 1970, Fig. 2B; Andrews et al., 1977, Pl. VI, Fig. 3.

Leclercqia complexa -- Kasper & Forbes, 1979, Pl. Fig. 1-12.

Incertae Sedis

- Prototaxites sp. -- Andrews & Kasper, 1970, p. 12.
Sciadophyton sp. -- Kasper et al., 1974, p. 358.
Thursophyton sp. -- Andrews et al., 1977, Pl. VI,
 Fig. 1-2.

Microfossils:

Spores (Andrews et al., 1977)

- Deltoidospora sp., cf. D. priddyi -- Pl. VII, Fig. 1.
Apiculiretusispora sp. -- Pl. VIII, Fig. 1-2.
Emphanisporites rotatus -- Pl. VII, Fig. 5; Pl. VIII,
 Fig. 5.
E. annulatus -- Pl. VII, Fig. 6-7.
 cf. Clivosisporites verrucata -- Pl. VII, Fig. 3.
Tholisporites sp., cf. T. chulus -- Pl. VII, Fig. 2.
Grandispora sp., cf. G. douglastownense -- Pl. VIII,
 Fig. 3-4.
Grandispora sp. -- Pl. VII, Fig. 4.

Chitinozoa (Andrews et al., 1977)

- Sphaerochitina sp. -- Pl. VIII, Fig. 6.

LOCALITY 1

Psilophyton dapsile
 ?Psilophyton sp.
Kaulangiophyton akantha
Taeniocrada sp.
Thursophyton sp.
Sciadophyton sp.

This locality was first reported by Dorf and Rankin in 1962 and has turned out to be the most productive fossil site in the Trout Valley Formation. The best-preserved plant here is the trimerophyte, Psilophyton dapsile (Kasper et al., 1974). P. dapsile was small, a few decimeters tall, with unornamented dichotomous or, occasionally, pseudomonopodial (overtopped) axes. The ultimate branchlets bore dense clusters of small sporangia. Specimens can be identified by their 2 mm wide axes, dichotomous branching and numerous elliptical paired sporangia about 2 mm long.

Also abundant at this site are specimens of a much larger plant tentatively assigned to the trimerophytes as ?Psilophyton

sp. (Andrews et al., 1977, Pl. II). Overtopping (pseudomonopodial habit) is quite evident in this plant with its distinct main axis and dichotomous laterals. The stems are 3 mm wide and are readily distinguished by the presence of longitudinal grooves and ridges probably resulting from supporting tissues revealed by compression. Specimens appear similar to those illustrated by Dorf and Rankin (1962, Fig. 6, 8) as Hostimella sp. (now: Hostinella) and Aphyllopteris sp. Hostinella and Aphyllopteris are both form genera--genera which are not readily assignable to any higher category because of the limited information provided by the specimens. The plant in question is referred to Psilophyton because of its overall habit. However, since sporangia have not been found as yet, a definite assignment to this genus cannot be made.

Kaulangiophyton akantha (Gensel et al., 1968) as reconstructed was a small plant with horizontal and erect axes up to 9 mm wide. Large (8 mm in diameter) ovoid sporangia on short stalks were borne attached along the erect axes. Small (2 mm long) spines were scattered along both prostrate and upright portions. K. akantha is important because of its intermediate position between the zosterophylls and the lycopods. It is tentatively classified here under the Zosterophyllophytina.

Taeniocrada (Andrews et al., 1977, Fig. 4) is a common Devonian plant genus described from around the world, however little is known about the plant itself. It is easily recognized by its broad (1½-2 cm) ribbon-like axes bearing a central strand 2 mm wide and dichotomizing infrequently. It is abundant at this locality.

Thursophyton (Andrews et al., 1977, Pl. VI, Fig. 1, 2) is another genus that is not known well, botanically speaking. In fact some authors use it as a form genus for fragmentary remains of axes densely covered with small, delicate, spine-like or hair-like enations. These 'leafy' axes branch pseudomonopodially or dichotomously, are about 5-6 mm wide and are clothed with emergences about 2 mm long. They are easily recognized but specimens are scarce. Sciadophyton has been found at this locality only once in the many years of collecting, so it too is rare.

LOCALITY 2

Pertica quadrifaria
Leclercqia complexa

This second locality was discovered by Forbes and I in July, 1971 (Kasper & Andrews, 1972); in later publications it is referred to as Locality # 7 (Andrews et al., 1977). Sev-

eral different plants are present at this site, two of which have been published: the trimerophyte, Pertica quadrifaria, and the lycopod, Leclercqia complexa.

As reconstructed, P. quadrifaria was a plant a meter or more tall with marked overtopping giving the appearance of a distinct main stem and side branches (Kasper & Andrews, 1972). The branches were arranged in a spiral and in four ranks or rows 90° apart (quadriseriate). The laterals were either fertile, i.e., sporangium-bearing, or sterile. Both types of branches were dichotomous and three dimensional. The sporangia are elliptical, 2-3 mm long and were aggregated into spherical masses. The specimens are easily identified by the large main stems (1½-2 cm wide) with dichotomously forking side branches to which are often attached the sporangial clusters.

At this locality but preserved in a very fragmentary manner is the lycopod, Leclercqia complexa (Kasper & Forbes, 1979). Identification of Devonian lycopods rests in large part on their leaf morphology. L. complexa is distinguished from other lycopods by its five-tipped leaves. The distal part of the blade is divided into a long tapering median segment with two shorter pointed segments on either side. Maceration of rock samples in HF acid frees large quantities of nearly complete leaves from the matrix. L. complexa can be recognized at the site by examining rock specimens with a hand lens. The highly coalified leaves are very reflective and readily show the five distal segments. L. complexa was first described from the late Middle Devonian Panther Mountain Formation of New York State (Banks et al., 1972). The implications of this as regards the age of the Trout Valley Formation are discussed in Kasper and Forbes (1979).

CONCLUSION

Although there is still much information to be obtained from the plant fossils of the formation, several important benefits have already accrued from studies to date. Botanically speaking, with the description of new genera and species the flora has provided a better understanding of the diversity of forms present during early Devonian times. Secondly, the remarkable preservation of some specimens--having large portions of the stem with fertile and sterile branch systems intact--has given us an opportunity to accurately reconstruct the plants in a true-to-life manner. Finally, the presence of several species of the genus, Psilophyton, displaying a wide variety of morphological features, has allowed us to speculate on possible evolutionary trends within the group (Kasper et al., 1974). It is a rather uncommon circumstance in early Devonian paleobotany to have several species of a single genus preserved within the limits of a geographically restricted formation.

Geologically speaking, accurate dating of the formation will probably rest on the plant megafossils. The spores, unfortunately, are poorly preserved "... showing a high degree of coalification ... as might have been expected from the high degree of diagenesis evident from the lithology" (Andrews et al., 1977, p. 275). Other paleontological data is scarce or yet to be discovered. Secondly, the morphological completeness of the plants of the Trout Valley flora permit identification and comparisons with isolated and fragmentary remains in other deposits in the region. At present, Forbes and I are working on two very small, geographically isolated and fragmentary floras in Nova Scotia. We are able to make identifications based on comparisons with the wider variety and more completely preserved plants of the Trout Valley Formation. Finally, it is hoped that after a thorough description, the flora of the Trout Valley Formation may serve as the basis along with other floras for a mega-plant biostratigraphic scheme for the region. Such a scheme would satisfy the initial request of the geologists working in the area who introduced us to the plant fossils with the hope that we could date the rocks and aid in correlating the numerous isolated intermontane continental deposits in the northern Appalachians. This is our ultimate goal and, at the same time, this demonstrates the important role to be played by the Trout Valley flora.

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ITINERARY

Assembly point is Shin Pond Lodge, Shin Pond, on State Route 159 about 10 miles northwest of Patten, Maine. Assembly time is 8:30 A.M. Topographic Maps: Traveler Mountain and Shin Pond Quadrangles.

Mileage

- 0.0 At Shin Pond Lodge take Grand Lake Road north out of Shin Pond (Shin Pond Quadrangle).
- 5.9 Cross bridge over the Seboeis River.
- 10.4 Note Forest Service Camp and side road (right) to Hay Lake.
- 14.7 Cross bridge over East Branch of Penobscot River (Traveler Mountain Quadrangle).
- 15.7 Baxter State Park entrance; Horse Mountain on left.
- 19.0 Pass through Trout Brook Farm area.
- 20.0 Road turns sharply left (south); enter Trout Valley Formation shortly.
- 22.9 Cross over Dry Brook.
- 23.4 LOCALITY 1: The Crossing--a picnic area on the right and before the bridge over Trout Brook; park cars here and take path along the southeast bank of Trout Brook for about 200 yards upstream to outcrop; refer to text for information on plant fossils.

AT NOON RETURN TO CARS and take South Branch Ponds road to South Branch Ponds Campsite (2.2 miles south of The Crossing) to eat lunch; AT 1:00 P.M. RETURN TO THE CROSSING and continue itinerary; cross bridge and continue on Grand Lake Road.

- 24.8 Cross 'first' Town Line between T6N, R9W and T5N, R9W.
- 25.7 Cross 'second' Town Line between T5N, R9W and T5N, R10W.
- 26.4. LOCALITY 2: Park along road, walk to Trout Brook about 75 yards south of road; large blocks of indurated bluish shale with maroon-colored impression/compression plant fossils are present in the stream bed; the origin of this material is the south bank of Trout Brook about 40 yards upstream.
- AT 4:00 P.M. RETURN TO CARS and drive back to Shin Pond Lodge.
- END OF TRIP.

FIGURES

Fig. 1. Psilophyton microspinosum. The pseudomonopodial habit, i.e., main stem and side branches, of this trimerophyte is evident. The side branches are either fertile (sporangium-bearing) or sterile. The sporangia are erect and borne in clusters. Note the small spines (=microspinosum). The scale bar is 1 cm.

Fig. 2. Psilophyton dapsile. This small trimerophyte is unornamented and branches dichotomously. The numerous sporangia are small, pendant, and in dense clusters. Plants were about 30 cm tall; scale bar is 1 cm.

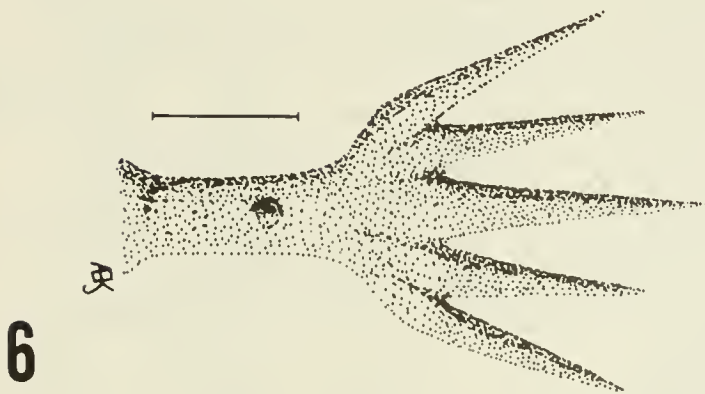
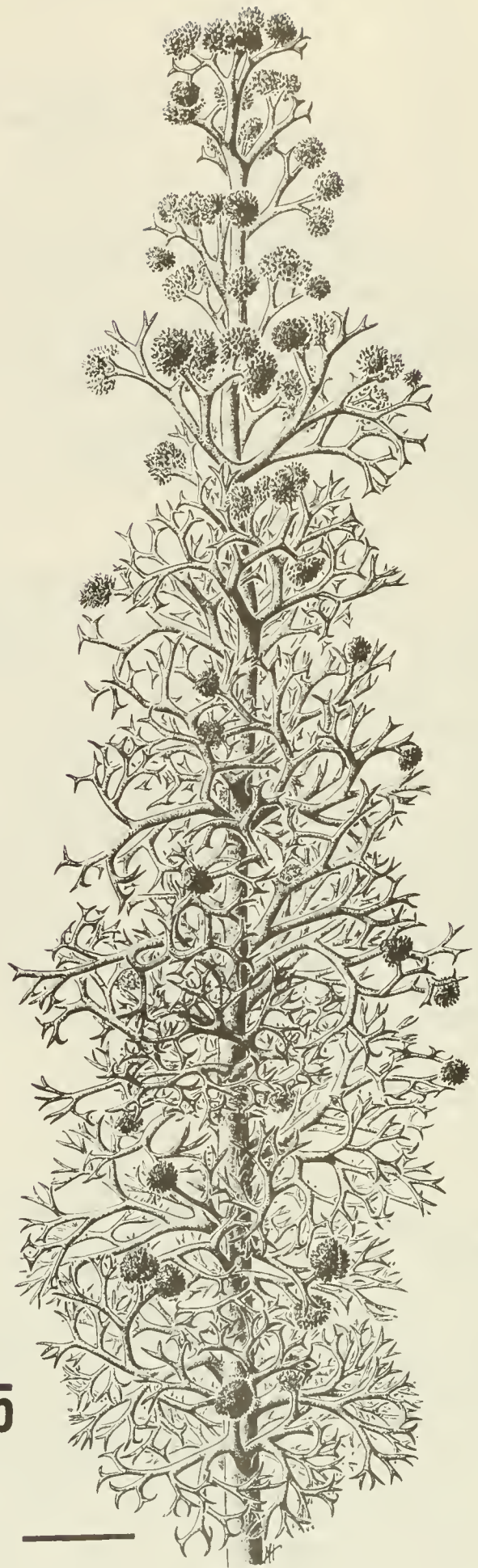
Fig. 3. Kaulangiophyton akantha. This plant as reconstructed has horizontal and erect axes, the latter bearing large sporangia on short stalks in lateral position--a zosterophyll characteristic. Because of its large axes and sporangia it approaches the lycopods in size. Scale is 2 cm.

Fig. 4. Psilophyton princeps. This historic trimerophyte shows pseudomonopodial branching, peg-like enations and large fusiform pendant sporangia in clusters on side branches. Major axes are about 1 cm wide. Reconstruction redrawn from Hueber (1967).

Fig. 5. Pertica quadrifaria. A trimerophyte about 1 meter tall with robust main stem and three-dimensionally branched sterile and fertile laterals. Fertile branches bear spherical masses of sporangia. Scale is 5 cm.

Fig. 6. Leclercqia complexa leaf. Stems of this lycopod were densely covered with small five-tipped leaves. The distal part of the blade was divided into a long median segment and two shorter segments on either side. The circular structure is the pad of tissue to which the single sporangium was attached--on the upper surface of the leaf. Scale is 1 mm.





ORDOVICIAN AND SILURIAN STRATIGRAPHY OF THE ASHLAND
SYNCLINORIUM AND ADJACENT TERRAIN

by

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The Ashland Synclinorium is a major northeast-trending tectonic feature in northeastern Maine. As shown in Figure 1, the synclinorium lies between the Pennington Mountain Anticlinorium to the northwest and the Aroostook-Matapedia Anticlinorium to the east. The synclinorium terminates with a southwest plunge in New Brunswick just northeast of the Siegas-VanBuren area (Hamilton-Smith, 1970) and may be traced for at least 60 miles (100 km) southwestward past Ashland, Maine. Between Ashland and the St. John River, Silurian rocks predominate in the synclinorium; southwest of Ashland Devonian rocks form the structure.

The tectono-stratigraphic significance of the synclinorium lies in its position between a region to the west that was uplifted during the Taconian Orogeny (essentially the Pennington Mountain Anticlinorium and terrain to the northwest) and an eastern area which was little affected by the orogeny and remained a deep-water (oceanic?) basin (essentially the terrain of the Aroostook-Matapedia Anticlinorium). As described elsewhere (Roy, this guide-book) the Silurian stratigraphy in the synclinorium is interpreted to be a post-Taconian continental margin sedimentary prism which may be "traced" southwestward at least to the rangely area of Maine. Here the sedimentary rocks of the prism are at low metamorphic grade (Richter and Roy, 1976) and rather simply deformed. The objective of this trip is to traverse the Ashland Synclinorium from Portage (on the Pennington Mountain Anticlinorium) to Presque Isle and examine pre-Taconian rocks and the Silurian lithofacies of the continental margin prism.

Stratigraphy

The stratigraphy of the region, including the synclinorium, has been described in some detail by Roy and Mencher (1976) and will be only briefly treated here. A more detailed map of the region covered by this trip and Trip C-5 is provided in Figure 2. STOPS 1 and 2 are in Ordovician rocks that were deformed during the Taconian and were subsequently the main source-rocks for the Silurian clastics. STOPS 3-9, 11, and 12 display the principal Silurian lithofacies. STOP 10 near Washburn shows the pre-Taconian "ribbon-rock" that is typical of the more eastern Aroostook-Matapedia Belt. For the purposes of describing the Silurian stratigraphy that we will examine it is convenient to divide the region into three subareas: I, the Frenchville area; II, the area of northward plunge of the Castle Hill Anticline; and III, the Mapleton-Presque-Isle area. Stratigraphic sections of these areas are given in Figures 3, 4, and 5. A fourth area in the vicinity of Jemtland, Maine, to the north, will be the focus of Trip C-5.

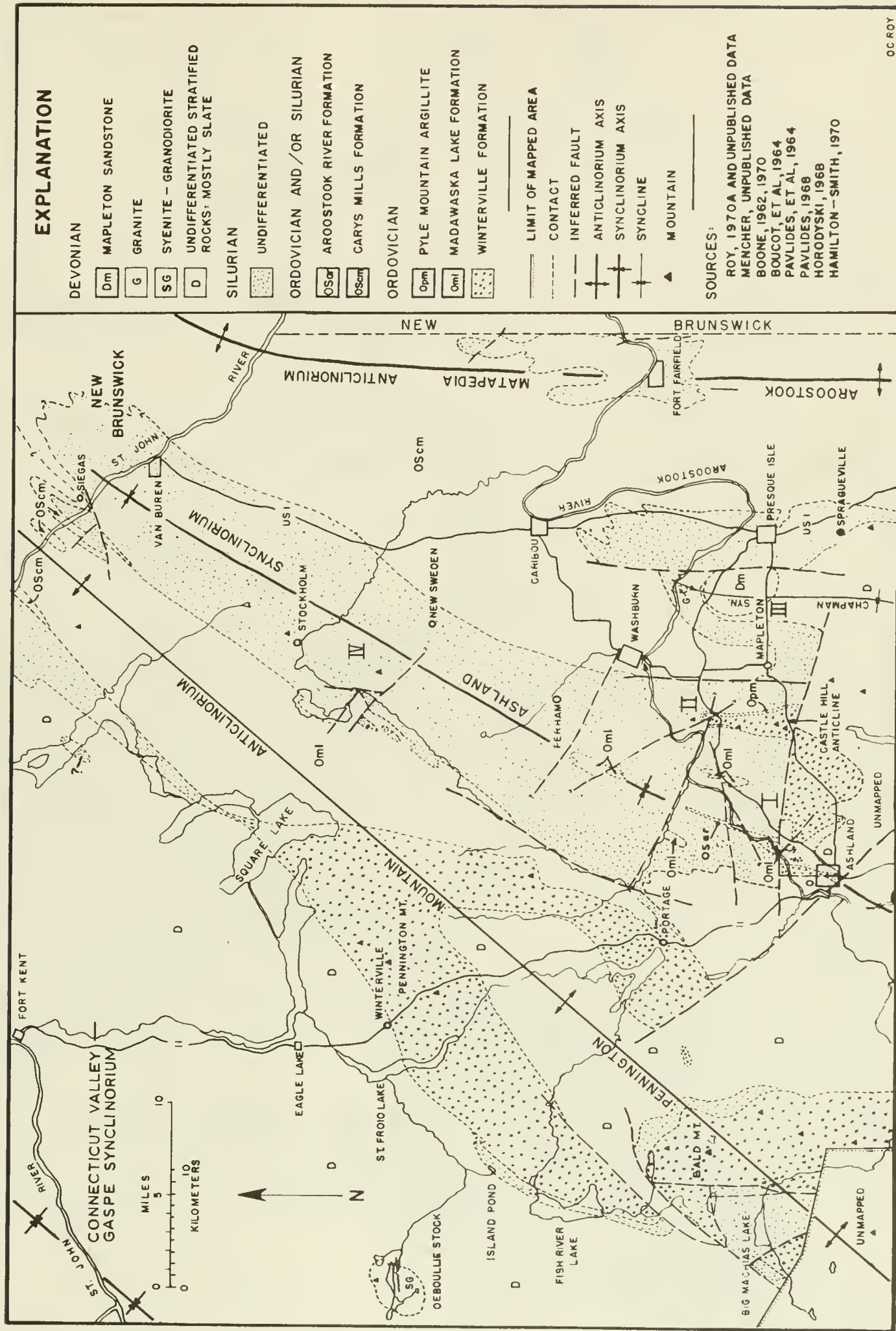


Figure 1: Generalized geologic and tectonic map of a portion of northeastern Aroostook County, Maine. (modified from Roy and Mencher, 1976)

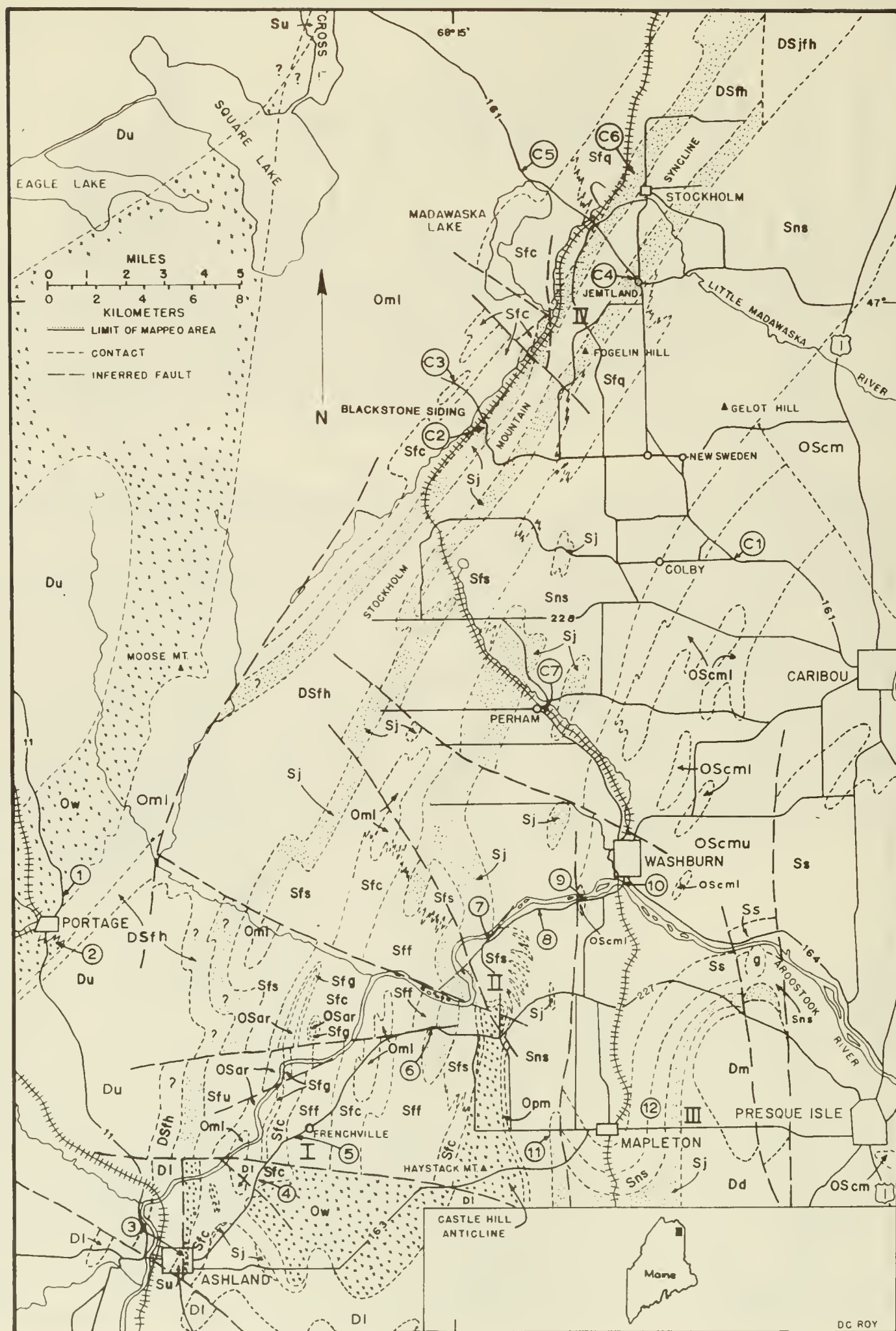
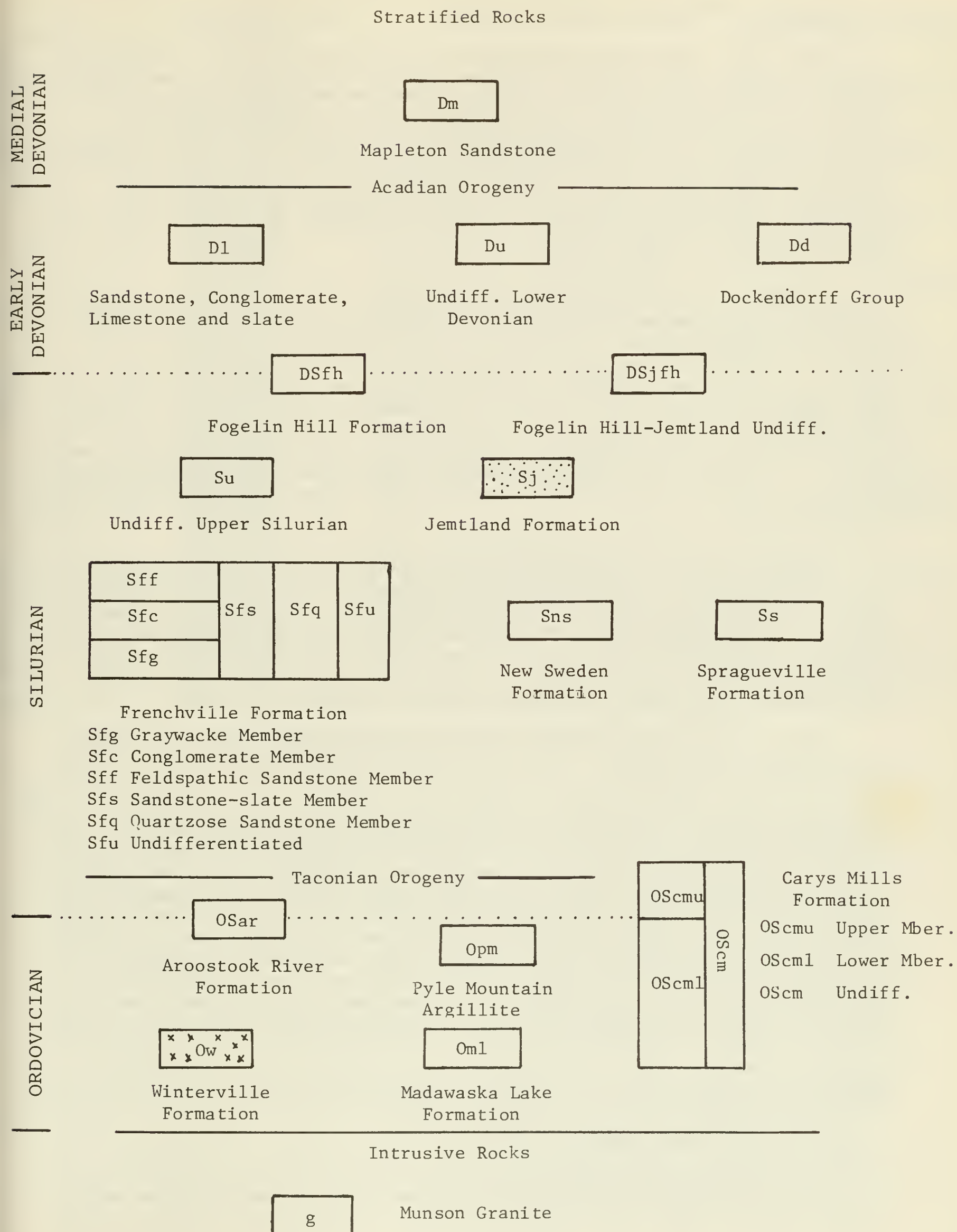


Figure 2: Geologic Map of the Ashland Synclinorium and adjacent areas. (modified from Roy and Mencher, 1976)

EXPLANATION FOR FIGURE 2



Ordovician Lithofacies

Three main lithofacies characterize the Ordovician System. In the west spilitic volcanic rocks, slate, graywacke and minor conglomerate comprise the Winterville Formation (STOP 1). Volcanic rocks predominate in exposures of the Winterville but sulfidic slate and graywacke are abundant in eastern portions of the formation and may be generally more important than indicated by outcrop frequency. Disseminated sulfides, especially pyrite, are ubiquitous in the Winterville and massive sulfide has recently been discovered in apparently commercial abundance near Bald Mountain west of Ashland. The Winterville is of Caradocian-Ashgillian age but the bottom of the formation is neither defined nor dated.

The Winterville Formation underlies much of the terrain in the vicinity of Ashland and in the core of the Castle Hill Anticline to the east (Figures 1 and 2). The extent of the Winterville in these areas is considerably south-east of its general surface extent farther north, but is similar to the eastern limit of rocks of similar age and lithofacies farther south (Pavrides, 1965, 1973). It is probable, but not as yet demonstrated, that the Ashland-Castle Hill Winterville connects in the subsurface with the main body of formation to the west. It is also possible that older phases (Caradocian and older) of the Winterville lithofacies extend beneath the Carys Mills Formation to the east and are continuous with similar rocks in the Tetagouche Group in New Brunswick.

In the vicinity of Portage (STOP 2) the Winterville interfingers with slate and graywacke of the Madawaska Lake Formation. The Madawaska Lake is also of Caradocian-Ashgillian age and is considered to be largely a temporal equivalent of the Winterville. The contact between the two formations between Portage and Square Lake (Figure 2) is considered to represent the zone of interfingering rather than a discrete plane of contact; it is possible that the portion of the contact from Moose Mountain northward is a fault.

Rocks of Late Ordovician age east of the Ashland Synclinorium are found in the Carys Mills Formation (STOPS 9 and 10; Pavrides, 1968; Roy and Mencher, 1976). In the Caribou-Washburn area the Carys Mills has been subdivided into two members (Roy, 1978b): the Lower Member is dominantly slate and graywacke with lesser micrite and contains all of the Ordovician fossil localities; the Upper Member is composed of slate and micrite in approximately equal proportions ("ribbon rock") with much lesser graywacke, and contains the Early Silurian fossil localities. This subdivision of the Carys Mills is not recognized in the type area by Pavrides (1976); although lithologically divisible in the eastern part of the region of Figure 2 it remains to be shown that the members as defined here can be carried more widely. The Carys Mills as mapped by Hamilton-Smith (1970; this volume) in the Siegas area of New Brunswick is considered here to be the Upper Member. In the Siegas area on the western flank of the Ashland Synclinorium, Hamilton-Smith has shown the "ribbon-rock" of the Carys Mills to overlies gradationally the Madawaska Lake Formation which suggests a westward onlap of the limestone-rich lithofacies. This onlap is concealed by younger rocks in the Ashland Synclinorium in the region of Figure 2.

Ordovician-Silurian Systemic Boundary

The transition from Ordovician to Silurian rocks in the region of Figure 2 is one of unconformity in the west and southwest and conformity in the east.

AGE		COLUMN (Thickness in Meters)	LITHOLOGY AND FEATURES		ENVIRONMENT	NOMENCLATURE	
EARLY DEVONIAN	Emsian						
	Siegenian						
	Gedinnian	?	760 +	Green and red silty shale and slate with interlayered thin beds of very calcareous, laminated fine-grained graywacke and siltstone.	Distal Flysch		Fogelin Hill Formation
SILURIAN	Pridolian	?					
	Ludlovian		760	Thinly interlayered gray shale, non-laminated silty shale, and fine-grained calcareous graywacke. Medium-grained and coarse-grained graywacke common. Turbidite structures typical of graywacke beds. Abundant graptolites.	Intermediate and Distal Flysch	Perham Group	Jemtland Formation
	Wenlockian						
			686	Medium-grained and coarse-grained feldspathic graywacke. Red shale with laminated ironstone present locally.			
	Llandoveryian	Upper					
		Middle	214	Medium-grained to thick-bedded polymictic conglomerate (75 to 80%) and lithic graywacke (20 to 25%) with minor pebbly and nonpebbly slate (1 to 2%).	Proximal "Coalesced Fans"	Frenchville Formation	Feldspathic Sandstone Member
		Lower	153				Congl. Member
ORDOVICIAN	Ashgillian	?		Gray and green-gray lithic graywacke.			
	Caradocian		610	Gray and green-gray slate thinly interlayered with calcareous fine-grained graywacke and siltstone. Medium-grained and coarse-grained graywacke common.	Intermediate and Distal Flysch		Aroostook River Formation
			610	Green-gray slate; minor graywacke.	Distal		Madawaska Lake Formation

Figure 3: Stratigraphic section for the Frenchville area (I).
(from Roy and Mencher, 1976)

Figure 6 shows the distribution of the angular unconformity together with the areas of inferred disconformable and conformable relations. Angular unconformities have been observed at two sites. In Ashland at STOP 1 and angular unconformity between sandstone and conglomerate of the Silurian Frenchville Formation and the underlying sulfidic slate and chert of the Ordovician Winterville Formation was observed during construction of a house. The unconformity is no longer visible but the Silurian rocks are well exposed and under favorable conditions small weathered exposures of the slate may be seen. The second place where the unconformity has been observed is near Blackstone Siding to the north (STOP 3 on Trip C-6) where Frenchville clastics overlies slate of the Madawaska Lake Formation.

The zone of disconformity shown in Figure 6 is inferred from the absence of marked angularity between Frenchville and pre-Frenchville units close to the contact and also the absence of earliest Silurian fossil localities in the lower Frenchville. Since the lower half of the Frenchville is not well dated the absence of earliest Silurian localities in that formation may be more of a sampling problem than an indication of hiatus. Therefore the area of disconformity in Figure 6 may best be viewed as a transitional zone between regions where unconformity are more confidently inferred.

AGE		COLUMN (Thickness in Meters)	LITHOLOGY AND FEATURES		ENVIRONMENT	NOMENCLATURE	
EARLY DEVONIAN	Emsian						
	Siegenian						
	Gedinnian						
SILURIAN	Pridolian	760	Thinly interlayered gray shale, non-laminated silty shale, laminated silty shale, and fine-grained calcareous graywacke. Medium-grained and coarse-grained graywacke common. Turbidite structures typical of graywacke beds. Abundant graptolites.		Intermediate and Distal Flysch	Perham Group	Jemtland Formation
	Ludlovian						
	Wenlockian	1070	Medium-grained to fine-grained graywacke with green-gray slate and minor micritic limestone. Graywacke beds display turbidite features. Pebbly graywacke and conglomerate locally near base.		Intermediate Coalesced Fans	Frenchville Formation	Sandstone and Slate Member
	Llandoveryian	Upper					
		Middle					
		Lower					
ORDOVICIAN	Ashgillian	90	Poorly cleaved silty argillite and green-gray slate. Silty argillite is commonly fossiliferous.		Distal Shallow Water		Pyle Mountain Argillite
	Caradocian	1500+	Andesitic, basaltic, keratophytic, and rhyolitic volcanic rocks and associated shallow intrusives; black slate; red, green, and gray chert; graywacke; and conglomerate.		Submarine Volcanism, and Sedimentation		Winterville Formation
			Disconformity or Slight Angular Unconformity		Submarine Erosion and/or Nondeposition		Taconian Hiatus

Figure 4: Stratigraphic section in the Castle Hill area (II).
(from Roy and Mencher, 1976)

The large region of systemic conformity is based on gradational upward transitions in shale-rich sections from dated Ordovician to dated earliest Silurian units. The sytemic boundary in most places is within the Carys Mills Formation and may be approximated by the transition from the lower to the upper member. The projection of an area of conformity toward Ashland along the Aroostook River is based on the presence in that area of the Aroostook River Formation which appears transitional between the Madawaska Lake and Frenchville formations and contains fauna of latest Ordovician or earliest Silurian affinities (Figure 3). North from Ashland a change from angular unconformity to probable conformity occurs over a very short distance. This distance may be shortened somewhat by the Alder Brook Fault located south of Frenchville (Figure 2), it is taken to be one indication of possibly steep post-Taconian submarine slopes.

Lower Silurian Lithofacies

The Lower Silurian contains three broad lithofacies: a western coarse-clastic facies (Frenchville Formation), a medial slate facies with lesser

AGE		COLUMN (Thickness in Meters)	LITHOLOGY AND FEATURES	ENVIRONMENT	NOMENCLATURE
EARLY DEVONIAN	Emsian			Volcanism and Marine Deposition	Dockendorff Group
	Siegenian			Erosion	Hiatus
	Gedinnian				
SILURIAN	Pridolian	760	Thinly interlayered gray shale, non-laminated silty shale, laminated silty shale and fine-grained calcareous graywacke. Medium-grained and coarse-grained graywacke common. Abundant graptolites.	Intermediate to Distal Flysch	Perham Group
	Ludovian	600	Gray, calcareous, phyllitic slate with lesser micritic limestone and minor graywacke, conglomerate, and green-gray slate. Lenticular bodies of laminated manganiferous ironstone with associated red and green slate. Graywacke and conglomerate more abundant in western part.	Distal Coalesced Fans	
	Wenlockian				Spragueville Formation
	Llandoveryian	1150	Thinly interbedded and interlaminated gray silty shale, light-gray silty limestone, and calcareous siltstone with lesser micritic limestone. Bioturbate structures are common.	Distal	
					Carys Mills Formation
ORDOVICIAN	Ashgillian	3800 ±	Thinly interlayered gray slate and argillaceous micritic limestone with lesser graywacke. Graywacke beds and many of the more silty and sandy limestone beds display turbidite features. Exposures have a "ribbon rock" appearance.	Distal Calcareous Flysch	
	Caradocian				

Figure 5: Stratigraphic section in the Mapleton-Presque Isle area (III).
(from Roy and Mencher, 1976)

micrite and ironstone (New Sweden Formation), and an eastern calcareous bioturbated mudstone facies (Spragueville Formation). The distribution of these lithofacies is shown in Figure 7. All three lithofacies contain fauna of Late Llandoveryian to Wenlockian age and the facies changes between them take place over fairly short distances. Facies changes across major folds within the synclinorium are common (e.g. in Castle Hill Anticline) and are thus best seen in the plunges of the folds; in some cases the facies changes take place in directions parallel to the structural grain (e.g. in folds west of Washburn).

The Frenchville Formation is the best studied unit (Roy 1970a, 1973) and is itself divided into five members. Three of the members have virtually no slate (Graywacke Member; Conglomerate Member; STOPS 3 and 4; and Feldspathic Sandstone Member, STOP 5) and consist of thick-bedded sandstone and conglomerate. These slate-poor members are vertically "stacked" in the Frenchville area (Figure 3). The Graywacked and Feldspathic Sandstone members are found only in

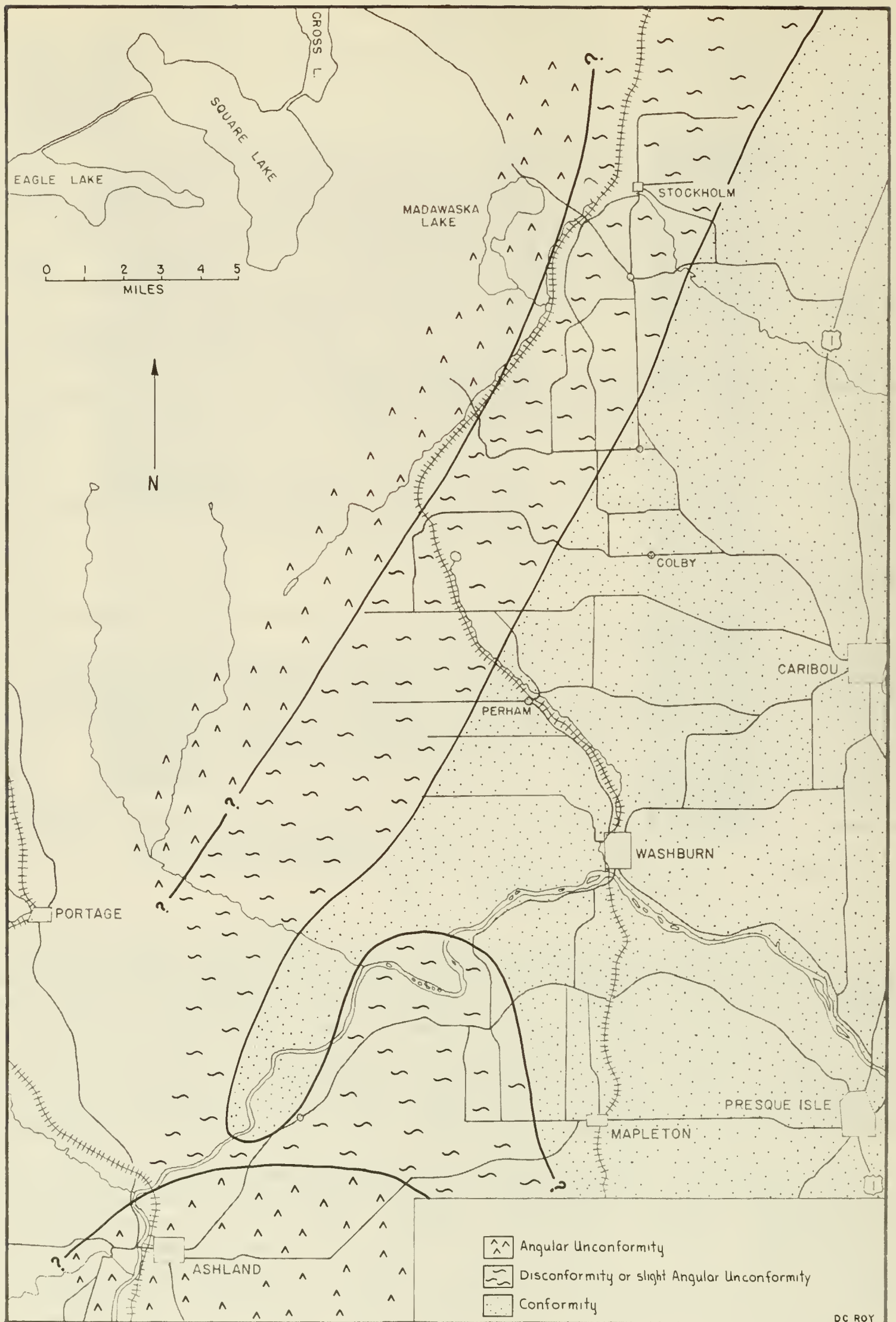


Figure 6: Nature of Ordovician-Silurian systemic boundary in the region shown in Figure 2.

the Frenchville area. The Conglomerate Member is much more extensive and may be mapped northward to the vicinity of Stockholm along the west flank of the Stockholm Mountain Syncline (Figures 2 and 7).

Two slate-rich members of the Frenchville lie between the slate-poor members and the eastern and more offshore lithofacies (Figure 7). The Sandstone-Slate Member (STOP 7) is the more southerly and differs from the Quartzose Sandstone Member in containing almost exclusively lithic and feldspathic graywacke. The Quartzose Sandstone Member is conspicuous in its abundance of quartz arenite and graywacke; this member is pretty much confined to the Stockholm area and represents a change in provenance rather than depositional environment.

The slate-rich Frenchville members represent the transitional facies from the base-of-slope deposits, represented by the slate-poor members, and the slate dominated facies farther east. The New Sweden Formation rests conformably on the Upper Member of the Carys Mills and differs from the older unit primarily in its preponderance of calcareous slate (STOP 8) and the local presence of laminated ironstone (STOP 11). The New Sweden is interpreted to be a generally deep-water deposit (Roy, 1970a). In the Mapleton area the Spragueville Formation (STOP 12) intervenes stratigraphically between the Carys Mills and New Sweden formations. Eastward the Spragueville thickens at the expense of the New Sweden Formation which it completely replaces at Presque Isle. The Spragueville lithofacies has been mapped as far east as Fort Fairfield by Pavlides (1968) and Roy and Mencher (unpublished data). The Spragueville consists of laminated, generally very calcareous, highly bioturbated mudstone which in the Presque Isle-Mapleton area is uncleaved. The terrigenous component of the Spragueville is coarser than that in the New Sweden and suggests that the source of the Spragueville material may not be from the west; alternate sources for the material have not been established but eastern sources seem likely. The bioturbation of the Spragueville facies complicates attempts to fit it with the non-bioturbated New Sweden and Frenchville lithofacies farther west. Bottom conditions during Spragueville deposition were apparently suitable to support an abundant but possibly not diverse infauna, whereas farther west bottom conditions were less supportive of sediment feeders. With time the bioturbated/non-bioturbated facies boundary moved eastward probably due to basin subsidence. The facies boundary does not appear to have moved east of Presque Isle before deposition of the overlying Jemtland Formation began.

Upper Silurian Lithofacies

The three Lower Silurian lithofacies are all succeeded gradationally by the distinctive thin-bedded flysch of the Jemtland Formation (STOP 6; Roy, 1970b). The bottom of the Jemtland Formation appears to be more or less synchronous everywhere and is of Early Wenlockian age (Roy and Mencher, 1976). The base of the formation is usually easy to determine in the field and has provided the closest thing to a structural marker horizon in the region. Many of the first order folds within the Ashland Synclinorium are therefore defined by the basal contact of the Jemtland (Figure 2). The source area for the formation was to the west and southwest as suggested by paleocurrent measurements and sandstone petrography. The petrology and sedimentology of the Jemtland is featured in Trip C-5 of this volume.

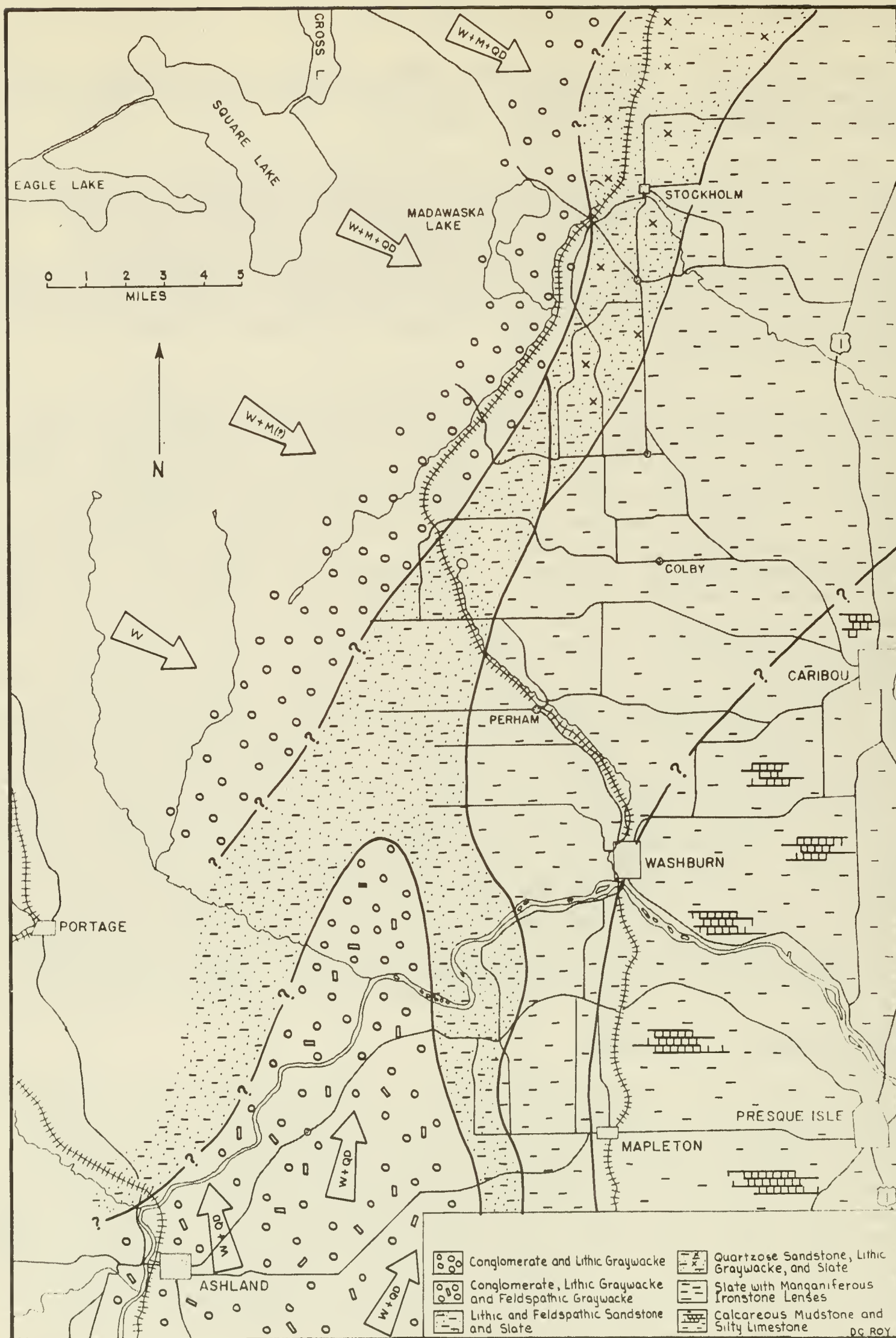


Figure 7: Major lithofacies of the Lower Silurian. Open arrows indicate principal source units: W, Winterville Formation; M, Madawaska Lake Formation; QD, Quartz Diorite.

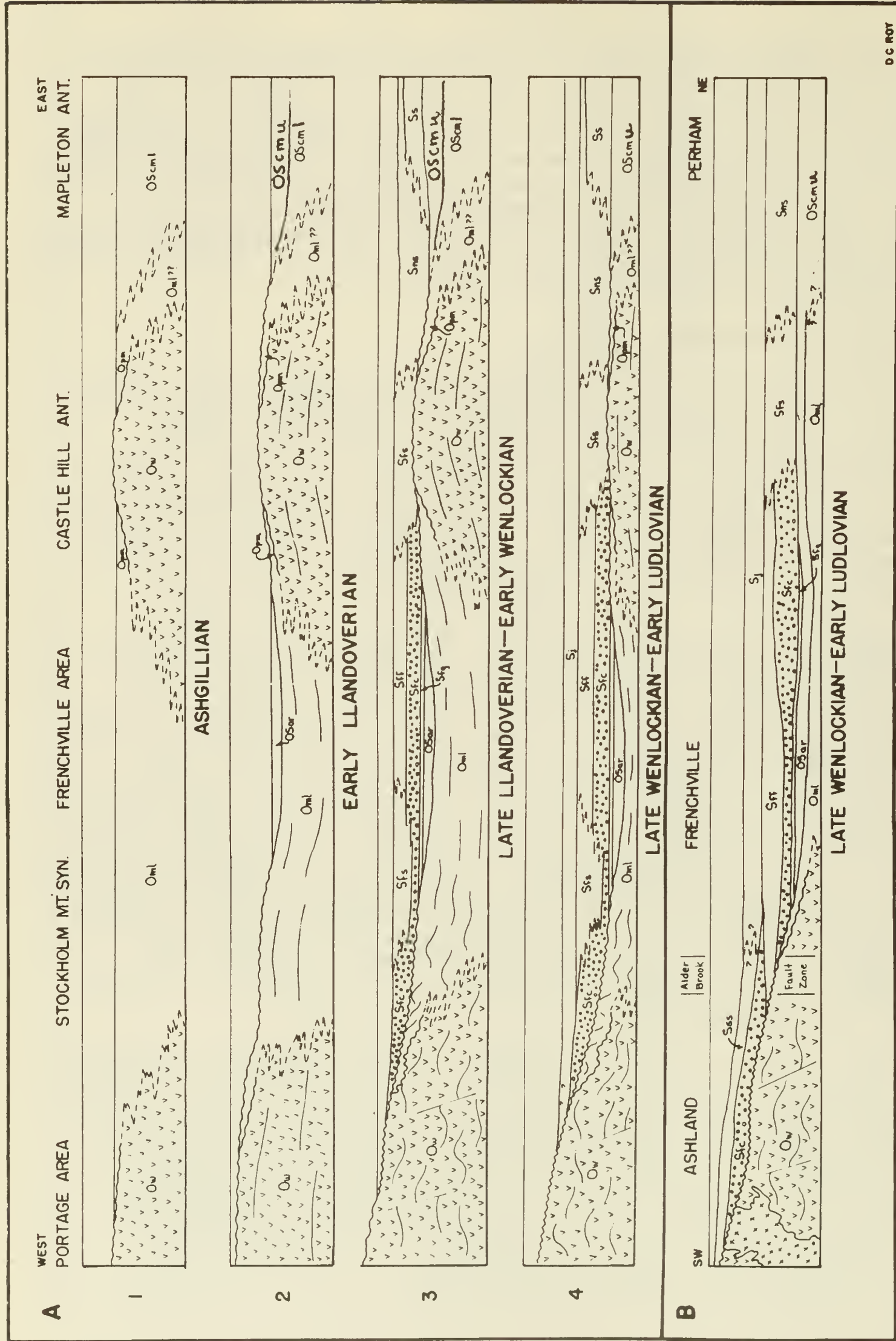
The Jemtland Formation is succeeded by the Fogelin Hill Formation in the Stockholm Mountain Syncline (see Figure 1 of Trip C-5). The Fogelin Hill Formation (not seen on this trip) is a distal flysch characterized by red and green slate interlayered with calcareous, laminated siltstone and fine-grained sandstone. The base of the formation is Early Ludlovian in age but it may range into the Early Devonian (Roy and Mencher, 1976). The Fogelin Hill is not present in the Presque Isle area; its absence there may be due to uplift associated with the Salinic Disturbance. The provenance of the Fogelin Hill is unclear but it is likely to have been derived from both western and eastern sources.

Tectono-Stratigraphic History

The history of Ordovician and Silurian (pre-Fogelin Hill) sedimentation in the region is summarized by the cross-sectional diagrams in Figure 8. Figure 8A shows four stages in the development of the stratigraphy along a cross-section from Portage to Mapleton. Figure 8B shows a similar cross-section from Ashland to Perham during the Late Silurian to illustrate the evolution of the Early Silurian clastic wedge derived from the south (see Figure 7) that was subsequently buried by sediments of the Jemtland Formation. The section of Figure 8B crosses each of those of Figure 8A approximately in the Frenchville Area.

The Taconian Orogeny may be dated by the earliest detritus deposited in the Ashland Synclinorium. Pavlides (1968) has suggested that the conformable transition from Carys Mills to younger units represents the first influx of sediment from Taconian uplands. This transition is approximately graptolite Zone 19 of the Llandovery. The turbidites of the Aroostook River Formation are inferred to also represent this initial detritus (Frame 2 of Figure 8A). The Taconian Orogeny produced uplift, mild folding, and subgreenschist grade alteration of the Winterville terrain in the western and southwestern parts of the area of Figure 2 (Richter and Roy, 1976). The Taconian effects in this region were late aspects of a longer period of deformation that closed an ocean basin between Portage and Quebec City (St. Julien and Hubert, 1975) and moved the continental margin of North America to the area of Figure 2 (Roy, 1976; this volume). Therefore the successive frames of Figure 8A illustrate the evolution of a Silurian continental margin clastic wedge.

The eastern margin of the Taconian upland was irregular during the Llandovery with a large embayment between Ashland and Portage (Figure 7). The distribution of Frenchville subfacies in the Frenchville area suggest a substantial coarse-debris source to the south. This source produced quartz diorite detritus (Feldspathic Sandstone Member) for a period of time. By Jemtland time the irregularity of the Taconian upland margin was somewhat reduced due to Late Silurian transgression onto the more subdued and submerging land area (Roy, this volume).



DC ROY

Figure 8: Cross-sections illustrating the evolution of the Ordovician and Silurian stratigraphy of the area shown in Figure 2. Upper horizontal line of each section represents sea level. Formations designated as in Figure 2.

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Itinerary

Mileage

- 0 Assembly point for the trip is in the parking lot adjacent and south of Judd's Store at the intersection of Routes 163 and 11 in Ashland. Starting time is 8:00 A.M. Drive north along Route 11 through "down-town" Ashland.
- .3 Intersection with Route 227 ("State Road") at the light in Ashland. Turn left (west) and continue along Route 11 toward Portage.
- 1.0 Bridge across the Aroostook River.
- 1.1 Intersection. Turn right and follow Route 11 north.
- 2.5 Bridge over Little Machias River.
- 2.6 Intersection with Wrightville Road (unpaved); continue north on Route 11.
- 3.15 B & A Railroad crossing.
- 6.9 Entrance to Pinkham's Mill. This mill is one of the largest saw mills east of the Mississippi and it is well worth the time to take the conducted tour.
- 11.0 Main intersection in Portage. Continue north on Route 11.
- 11.8 STOP 1. Park in turn-out at the edge of the golf course. Exposure is part of an old road metal pit that has been somewhat "groomed".

Black sulfidic slate and volcanic rocks of the Winterville Formation are exposed here in the Portage Anticline of the Pennington Mountain Anticlinorium. Cleavage in the slate is parallel to bedding and graptolites are common. W.B.N. Berry assigns the graptolites to Zone 12 (Climacograptus bicornis Zone) of the Ordovician. It is unclear at this locality whether the basaltic rocks are intrusive or extrusive. The volcanic rocks here and elsewhere in the Winterville Formation have not been studied in detail. These rocks are inferred to have been deformed during both the Taconian and Acadian orogenies. They are presently at Prehnite-Pumpellyite grade. On a clear day there is a nice view southeastward across the Ashland Synclinorium (low terrain) toward the Castle Hill Anticline with Haystack Mountain on the south end of the axial ridge which is cored by Winterville Volcanic rocks.

Turn vehicles around and head south on Route 11.

12.6 Main intersection in Portage. Continue south on Route 11.

12.85 Turn left (east) onto small gravel road. Access road appears to be part of a driveway for a small green house but continues beyond the house, along the edge of a field, and into the woods.

STOP 2. Park in the road metal pit at the end of the access road.

This pit, and another to the north, expose typical slate of the Madawaska Lake Formation. These rocks are interpreted to interfinger southwestward with volcanic rocks of the Winterville Formation. The slate is typically gray to olive-green in color, fracture cleaved, and very fine-grained. Orange weathering, laminated and cross-laminated fine-grained sandstone or siltstone are usually present as thin beds. More massive dark gray lithic graywacke is locally abundant. Here layers of volcanoclastic sandstone and lithic tuff are included in the Madawaska Lake. Along the west wall of the pit is an apparently graded lithic tuff bed with pumice/scoria fragments near its top and in the overlying slate.

Return to Route 11 and head south to Ashland.

23.8 Bridge across the Aroostook River.

24.5 Intersection of Routes 11 and 227 in Ashland at stop light; continue straight on Route 227 heading east.

25.0 Left (north) on Cottage Hill Road near water tank.

25.2 STOP 3. Park on the right side of the road leaving spaces for the driveways. The outcrops for this stop are on the property of Mr. and Mrs. Judson Holmquist.

During the construction of the Holmquist house the Taconian unconformity was exposed in the east wall of the basement excavation as depicted in Figure 9. The unconformity (now covered) is between sulfidic slates and lesser chert of the Ordovician Winterville Formation and the overlying Conglomerate Member of the Silurian

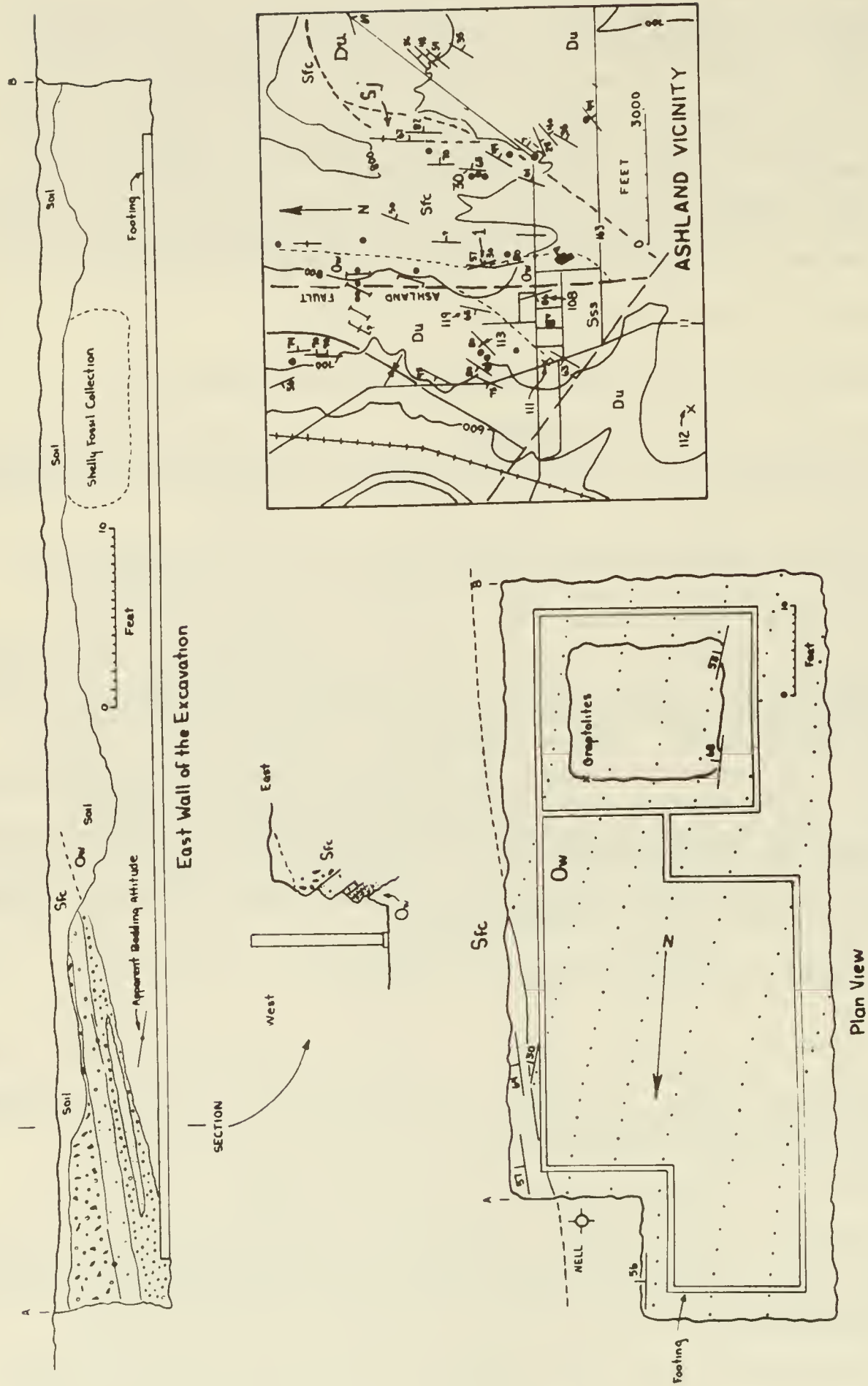


Figure 9: Geologic maps of Ashland and the basement excavation for the Holmquist house (built and formerly owned by Mr. and Mrs. Carlton Jimmo). Fossil localities are indicated on the Ashland map; STOP 3 is at fossil locality 1. Formations designated as in Figure 2.

Frenchville Formation. Both shelly and graptolite fauna were recovered from the Ordovician slate. The graptolites suggest to W.B.N. Berry (Roy, 1970a) an age within the span of zones 12 to 15 of the Ordovician; the shelly fossils have been provisionally assigned to the Ashgillian of the Ordovician by Neuman (1968, p44). No dateable fauna have been extracted from the Frenchville here but a shelly collection from Locality 30 (Figure 9) is of Late Llandoveryan-Wenlockian age according to A.J. Boucot (Roy, 1970a). The Frenchville conglomerate here is polymictic with provenance dominated by the Winterville Formation. Lenses of lithic graywacke and arenite are common. Limestone clasts and shell debris are more common here than elsewhere in the conglomerate member and form part of the evidence for the conclusion that the member is here of more shallow-water origin than farther north. A major north-south fault is present just west of the house, at or near the base of the escarpment (Figure 9). In the lowlands to the west of the fault, one finds only Late Silurian and Early Devonian units represented.

We will make a U-turn (carefully) and return to Route 227.

- 25.4 Turn left onto Route 227 and continue east.
- 25.85 Poor exposures of Lower Devonian sandstone at curve in road. The Devonian rests on Silurian and Ordovician rocks unconformably in the Ashland area. The unconformity represents the Salinic hiatus. The lower most Devonian contains abundant limestone, lithic sandstone, polymictic conglomerate, and monomictic mudstone conglomerate (with clasts of the Jemmland Formation or volcanic rocks).
- 27.65 Roadcut exposure on the right (east) of basalt of the Winterville Formation.
- 28.60 STOP 4. Park on road shoulder and in a small "turn-out" on the east side of road. Walk east across small field to woods line; outcrop is a few tens of feet in woods. We are here on the property of Mrs. Mabel Berry.

This is the best easy-access exposure of conglomerate of the Frenchville Formation. These beds are a few hundred feet stratigraphically above the presumed unconformity between the Frenchville and Winterville formations. These massive roundstone pebble and cobble conglomerates are thought to represent material transported down a submarine slope into relatively deep water. The conglomerates show little or no evidence of current sorting and commonly show a more or less "open" texture suggestive of rapid sedimentation. Lesser lithic arenite is also present here.

Returns to cars and continue north on Route 227.

- 28.75 Road ditch exposures on the right (east) near and down hill from the power pole consist of pebbly mudstone of the Frenchville that is stratigraphically above the beds at STOP 4. These exposures have yielded Late Llandoveryan brachiopods.

- 29.05 Bridge across Alder Brook. We here cross the Alder Brook Fault which is a major east-west fault that extends eastward well into the Presque Isle Quadrangle.
- 29.90 STOP 5. Park in turn-out just west of large road cut on right (south) side of highway. Walk to exposure.

This is the "type" exposure of the Frenchville Formation as established by Boucot and others (1964). The exposure is located below the site of the former Frenchville church and lies on the west flank of the broad Frenchville Syncline (Figure 10). These sandstones are typical of the Feldspathic Sandstone Member of the Frenchville. Notice the near absence of current-produced structures, homogeneous distribution of lithic and fossil fragments, massive bedding, and apparent graded bedding in some of the massive beds. Figure 11 gives a generalized stratigraphic section for this outcrop. The sandstones contain brachiopods from three of Ziegler's (1965) depth communities (Eocoelia, Pentamerus, and Stricklandia) which are taken to indicate downslope transport of both sand and shell material to at least the depth of the deepest community (Stricklandia; 30-60 meters). The Winterville does not appear to be the major source for the detritus here; it is inferred that a Taconian quartz-diorite like the Rockabema Quartz Diorite near Shin Pond to the south was the source of the material.

Return to cars and continue north on Route 227.

- 32.7 Castle Hill Townline.
- 34.6 Small store on left (north) side of road. Usually soda and munchies are available.
- 34.9 Demarchant Brook.
- 35.0 STOP 6. Park in grassy field at top of the hill on the right (south) side of road. This may also serve nicely as a LUNCH STOP.

This is the best outcrop of the Jemtland Formation that we will be near today. The sedimentology of this formation is the focus of Trip C-5 where larger exposures in its type area will be examined. This exposure is typical of the formation both in its lithology and in the abundance of graptolites which Berry (1960) has fully described. These beds are assigned to zone 33 of the Silurian (Early Ludlovian). Note the three types of shale, preservation of delicate silt laminae, ripple cross-laminae and locally preserved ripples, micro load-cast features associated with the siltstone laminae, and convolute bedding. Rusty orange weathering (typical of the formation) is due to oxidation of iron in pyrite and ankerite (?). This formation overlies the Frenchville Formation gradationally and is interpreted to represent deposition in deep, relatively quiet water.

Return to cars and continue east on Route 227.

- 35.25 From the crest of the hill here we have a nice view eastward of a line of hills that marks the core of the Castle Hill Anticline. From south

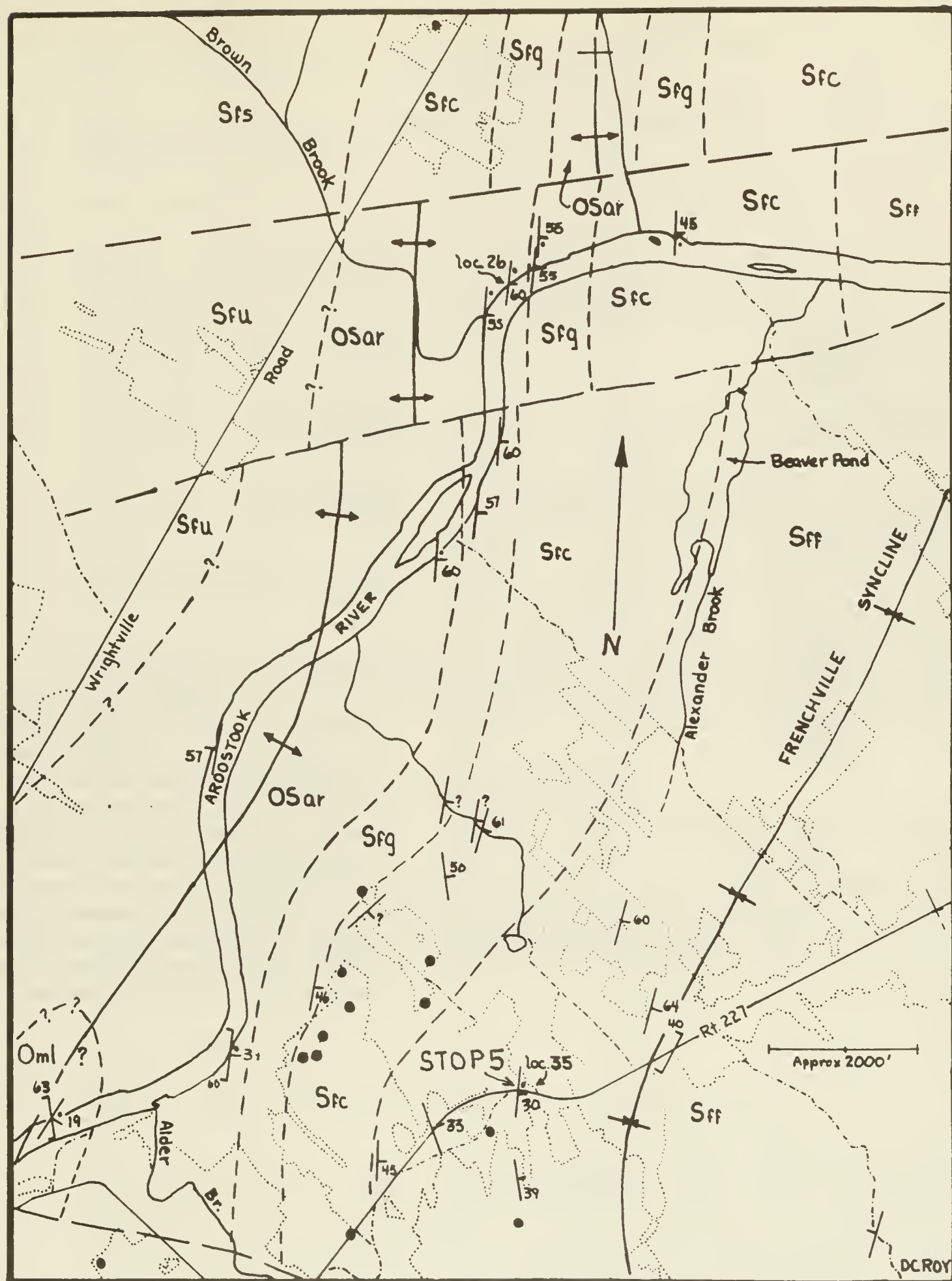


Figure 10: Geologic map of Frenchville area showing the location of STOP 5. Dotted lines outline cleared fields; base map uncorrected airphoto. Formations are designated as in Figure 2.

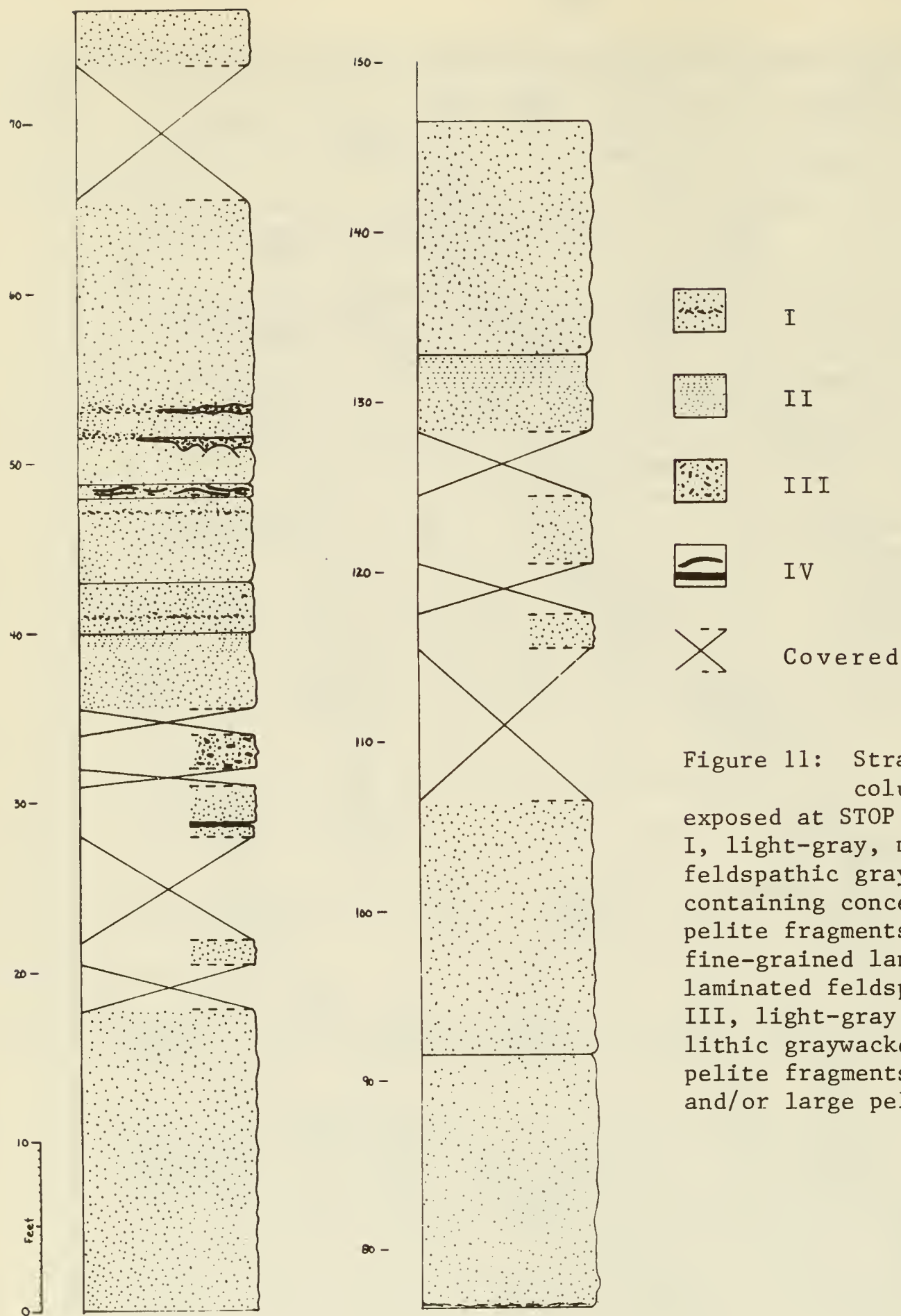


Figure 11: Stratigraphic column for section exposed at STOP 5. Rock types: I, light-gray, medium-grained feldspathic graywacke locally containing concentrations of pelite fragments; II, light-gray, fine-grained laminated and non-laminated feldspathic graywacke; III, light-gray, coarse-grained lithic graywacke with abundant pelite fragments; IV, shale beds and/or large pelite fragments.

to north the hills are Haystack Mountain (conical in shape), Pyle Mountain, McDonald Mountain and Castle Hill. Our next stop is in the area of northward plunge of the Castle Hill Anticline.

- 36.7 Turn left (north) just beyond small church.
- 36.85 Turn left (north) onto gravel road. This road takes us more or less along the crest of the Castle Hill Anticline.
- 38.35 Steep descent here is down-dip in the sandstone-slate member of the Frenchville Formation.
- 38.95 Wade Townline.
- 40.3 STOP 7. Park on grassy turn-out on left (north) side of road at sharp bend and walk through field northward to the Aroostook River. Carefully descend to the river bank and walk upstream about 1000 feet to the large exposure (see Figure 12).

The sequence exposed here (Figure 13) is transitional between the slate-poor sandstone/conglomerate western phase of the Frenchville and the slate dominated New Sweden Formation to the east. This sandy-flysch phase of the Frenchville has been separated out as the Sandstone/slate Member and is interpreted to represent an intermediate-to-distal submarine fan sequence. The graywacke beds show many of the classic features of turbidites and with care more or less complete turbidites may be sampled. This is fossil locality 43 of Roy and Mencher (1976) and it has produced conodonts which Gil Klapper places in the interval of C_2 - C_4 of the Late Llandoveryan and abundant *Artriotreta* brachiopods studied early by R.B. Neuman and more recently by W.H. Forbes who has been extensively "mining" the limestone beds for fossils since he and I made the first collections in the late 1960's.

Return to the cars and continue eastward.

- 41.3 STOP 8. Park on shoulder of road. Outcrop on right side of road.

The low exposures show typical phyllitic slate of the New Sweden Formation which is the eastern equivalent of the Frenchville we have been looking at. Micritic limestone beds, locally lenticular due to bedding transposition by cleavage, are generally present in the formation.

Continue eastward.

- 42.5 STOP 9. Park carefully on the shoulder of the road.

This large exposure shows the slate and graywacke typical of the Lower Member of the Cary's Mills as I have mapped it in the Caribou Quadrangle. This section faces eastward and gives way vertically to the "ribbon rock" Upper Member of Cary's Mills seen in exposures on the river bank just east of the outcrop here. Graptolites of Late Ordovician (Ashgillian) age were reported from here prior to my work and confirmed by W.B.N. Berry at my request. I have not been able to duplicate the collection, however, and would appreciate any finds you can make. Graptolites of Zone 18 of the Llandoveryan have been found at the "ribbon rock" exposure on the river just down-stream from here, thus indicating that these beds are Zone 18 or older.

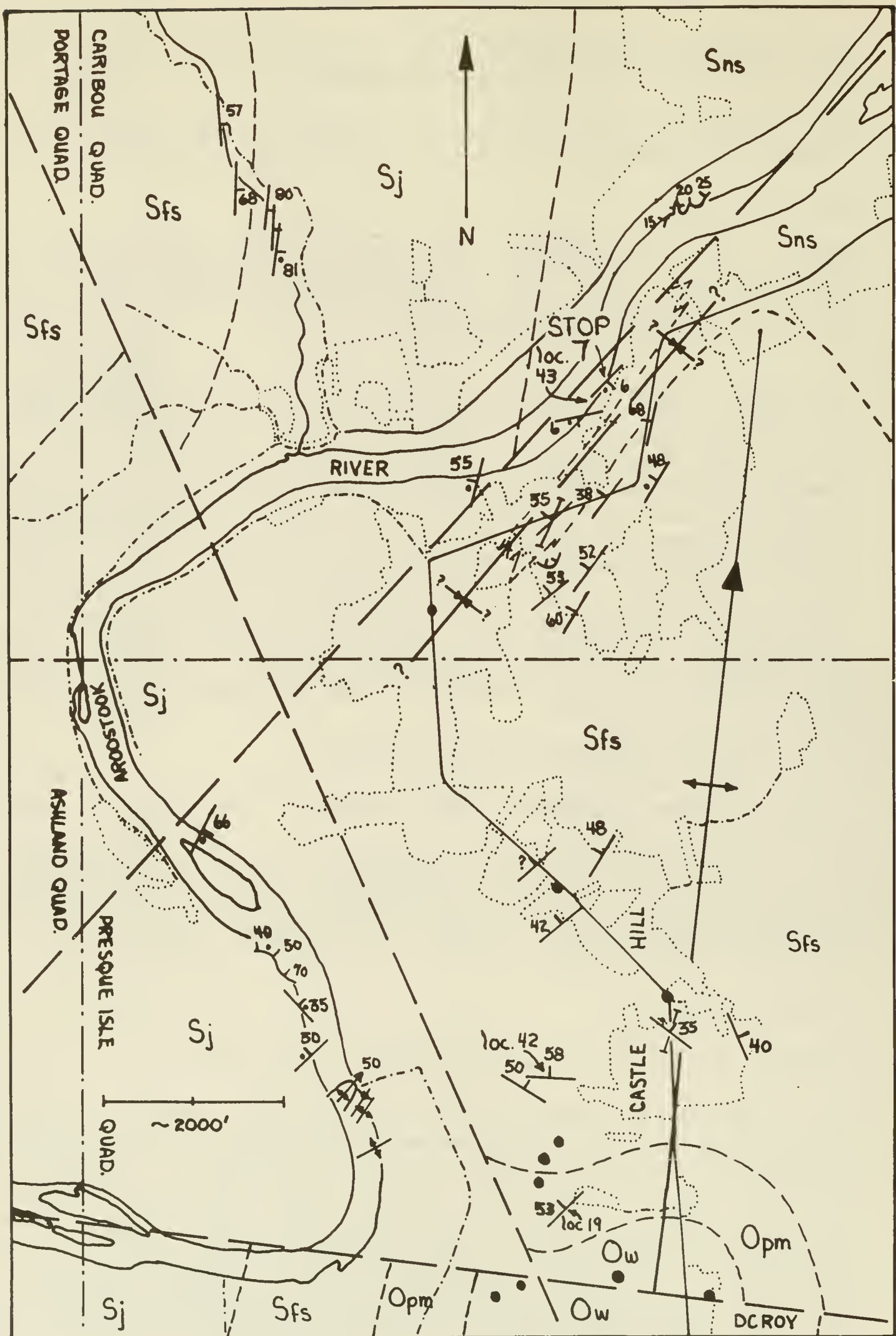


Figure 12: Geologic map of the Castle Hill area showing the location of STOP 7. Dotted lines outline cleared areas; base is on uncorrected airphoto. Formations are designated as in Figure 2.

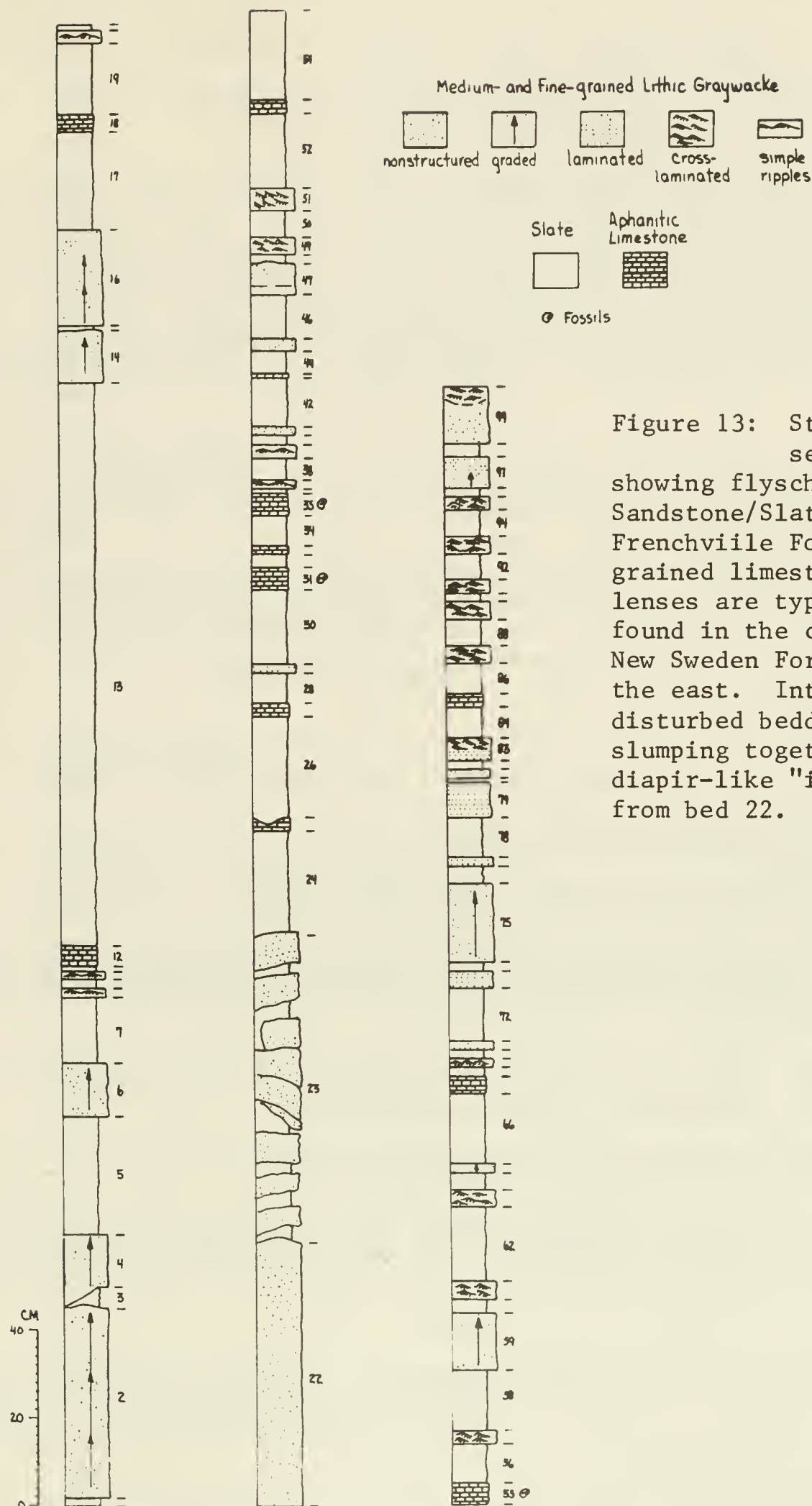


Figure 13: Stratigraphic section at STOP 7 showing flysch character of the Sandstone/Slate Member of the Frenchville Formation. Fine-grained limestone beds and lenses are typical of those found in the contemporaneous New Sweden Formation nearby to the east. Interval 23 shows disturbed bedding caused by slumping together with upward diapir-like "intrusion" by sand from bed 22.

Return to cars and continue eastward.

- 43.15 Intersection. Continue straight (to the east).
- 45.70 Bridge over the Aroostook River at Bugbee just south of Washburn.
- 45.80 Turn left just east of bridge and park after making a U-turn. Descend to the river and exposures along the east bank.

STOP 10. These exposures display typical "ribbon rock" limestone and slate of the Upper Member of the Cary's Mills. All of the Silurian fossil localities in the Cary's Mills of this region come from the Upper Member of the formation and from rocks such as these. Micritic Limestone is generally at least 50% of the unit; the slate is typically medium-to-dark gray and quite calcareous. The limestones are generally impure with at least 20% or so argillaceous minerals and quartz. The impurities may be graded but are more commonly concentrated along "mica-parting" laminations and cross-laminations which show that currents were important in the deposition of the limestone beds. It is likely that most of these limestone beds are deep-water calcisiltites deposited by turbidity currents as suggested by Pavlides (1965, 1968); some may be "pelagic" oozes. The more rust-weathering beds are probably ankeritic as suggested by Pavlides and possibly more pyritiferous. Detailed studies of this "ribbon rock" have not as yet been done, but are important (and made difficult) when the lateral extent of this lithofacies is considered.

Return to cars.

- 45.9 Intersection at east end of bridge; turn right and head west.
- 46.5 Bear left (south) at intersection.
- 49.45 Intersection with Route 227. Turn right (west).
- 50.30 Exposure on left (south) of basal Jemtland Formation preserved in a tight syncline and underlain by the New Sweden Formation. View west of Castle Hill.
- 52.20 Turn left (south) on Turner Road. Turner road parallels the axis of Castle Hill Anticline along its east flank.
- 53.20 Good view of Pyle Mountain to the right front.
- 54.15 Small pit on the left (east) side of road contains red slate of the lower most New Sweden Formation. A thin calcarenite lens at this locality produced a Late Llandoveryian fauna. The lens has now been more-or-less completely removed.
- 54.60 Intersection of Turner and Dudley roads. At this intersection stood the Pyle School near which Ashgillian trilobites have been recovered from an argillite which is now named after Pyle Mountain. Turn left (east) on Dudley Road.
- 55.65 Turn right onto farm road and drive into a small road-metal pit. We are here the guests of Mr. and Mrs. Delance Lovely.

STOP 11. This relatively fresh pit exposes red and green slate of the New Sweden Formation and a basal mudstone conglomerate of the Jemtland Formation. An approximately six-foot interval of laminated, mangani-ferous ironstone, is exposed within the New Sweden. Such ironstone deposits are widespread both areally and stratigraphically within the formation and collectively form the so-called Aroostook manganese deposits. The Dudley deposit, described as one of the richest in manganese by Miller (1947), is along strike from this pit to the north across Dudley Road. The red and lesser green slates are common in the vicinity of the thicker laminated ironstones and are undoubtedly genetically related. The varigated slate replaces the calcareous phyllitic slates (STOP 8) that characterize most of the formation. The basal mudstone conglomerate contains fossiliferous limestone cobbles and pebbles; near here the conglomerate has yielded Late Silurian (Ludlovian) brachiopods.

Return to Dudley Road.

- 55.85 Intersection of farm road and Dudley Road; turn right and head eastward.
- 56.80 Intersection of Dudley Road and Route 163. Continue eastward on Route 163 toward Mapleton.
- 57.40 Railroad overpass in Mapleton.
- 58.50 STOP 12. Park on shoulder of highway; exposures on the right (south) side of highway.

This is a typical exposure of the Spragueville Formation which is the eastern-most temporal equivalent of the New Sweden and Frenchville formations. The Spragueville consists of laminated highly calcareous mudstone. Light-gray laminae are typically 50% or more carbonate with quartz and muscovite/chlorite being the principle insoluble material. Microfossils are common in these calcareous laminae. This formation is extensively bioturbated with abundant interstratal and intrastratal burrows. Bioturbation locally completely disrupts lamination but generally the original laminated character of the deposit is clearly evident. Here the Spragueville underlies the New Sweden Formation which thins to extinction near Presque Isle where the Spragueville lies conformably between the Cary's Mills and Jemtland formations. The Spragueville lithofacies extends into Canada east of Fort Fairfield but has not everywhere been separated from the Cary's Mills. Also of interest here is a poorly exposed dike of teschenite, an analcime-rich diabase.

This is the last stop. Continue east to Presque Isle on Route 163.

- 59.70 Cuts of Early Devonian volcanic rocks; part of the Hedgehog Formation of the Dockendorf Group.
- 61.80 Road cut on the left (north) side of the highway of the Middle Devonian Mapleton Formation. These conglomerates are part of the

oldest clearly post-Acadian molasses in the northern Appalachians. The Mapleton rests with a nearly 90 degree angular unconformity on rocks of the Acadian-deformed Dockendorf Group and contains clasts of most of the pre-Acadian units in region.

- 64.2 Intersection with Pleasant Street in Presque Isle. Turn right.
- 64.6 Intersection of Route 163 and U.S. 1 in downtown Presque Isle. Turn right (south) and find your own way from here on.

TRIPS B-7 and C-2

BEDROCK GEOLOGY OF THE PRESQUE ISLE AREA

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INTRODUCTION

This trip includes stops in the Presque Isle, Caribou, and Ashland 15-minute quadrangles. The author mapped the north-central part of the Presque Isle quadrangle in 1960 as an undergraduate thesis at the Massachusetts Institute of Technology and returned in the summers of 1961 and 1962 to extend reconnaissance bedrock mapping to the west and north.* Although maintaining an active interest in the area and visiting it annually, the author must stress that nearly twenty years have passed since he last mapped in this area.

The bedrock geology of the Presque Isle Quadrangle was published by Boucot, Field, Fletcher, Forbes, Naylor, and Pavlides (1964). That paper outlines the previous history of geologic mapping locally. Briefly, the area has long been known for its rich and abundant fossil localities and many notable paleontologists have published articles about the area. Bedrock mapping lagged however. The pioneering work of H.E. Gregory (Williams and Gregory, 1960) produced a useful lithologic map, but the work was not stratigraphically or structurally oriented. Geologists from the Maine and United States Geological Surveys mapped parts of the area in the 1940's but were concerned chiefly with the so-called Aroostook County manganese deposits and their immediate geologic setting. The project on which the author and others worked in the early 1960's under the supervision of A.J. Boucot was thus the first general purpose stratigraphic and structural mapping of this interesting area.

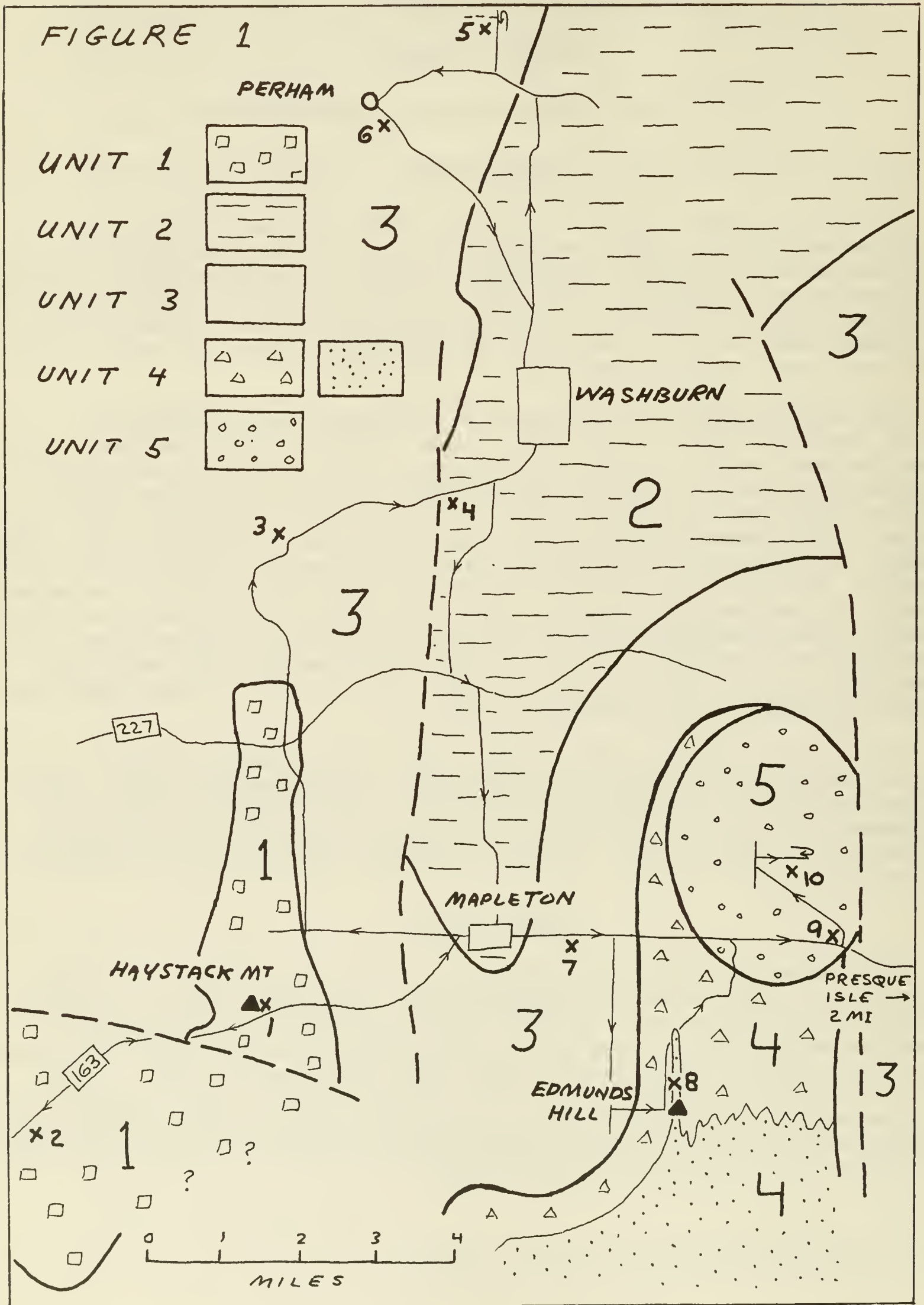
D.C. Roy, Ely Mencher, and co-workers began their extensive mapping project immediately after the end of the phase of work described above. Their results, mostly in areas to the north and west of the Presque Isle quadrangle are summarized in the introduction to this Guidebook and in the articles by Roy and are not repeated here. Their stratigraphic nomenclature is followed in this article.

GENERAL GEOLOGY

The Presque Isle quadrangle exposes four major litho-tectonic rock units that extend to other parts of the Appalachians. In no other area are all four units so closely juxtaposed. Interest is further heightened by the richly fossiliferous character of most of the units in this area, providing age assignments that can be usefully extrapolated to other less-fossiliferous areas. A fifth unit, the Mapleton Formation, appears to be unique to the Presque Isle area, and is indispensable in pinpointing the age of the early phase of the Acadian Orogeny. The five groups are outlined below, and are shown in Figure 1 (adapted from Roy and Mencher, 1976; Boucot, and others, 1964).

*Work supported by the Maine Geological Survey.

Geology of the Presque Isle area showing lithotectonic units described in text, with field trip route and stops.



Unit 1 comprises volcanic and sedimentary rocks exposed in the Castle Hill Anticline in the northwest corner of the Presque Isle quadrangle and in the Ashland quadrangle to the west (stops 1 and 2). The volcanic rocks are a bimodal suite comprised mainly of greenstone (spilite) and sodium-rich felsite (Keratophyre). These are closely associated with sulfidic, graptolite-bearing shale, red or black chert, and green argillite. These rocks are of Middle and Upper Ordovician age. This belt of rocks extends (with some gaps) for hundreds of kilometers to the southwest and probably links with the Ordovician metavolcanic and metasedimentary rocks of the Bronson Hill Anticlinorium. Exposures die out only a few kilometers to the north, but the belt is postulated to continue some distance further north below the surface (Hamilton-Smith, 1970). The author and others have speculated that this belt is one of perhaps half a dozen volcanic arcs significant in the evolution of the Northern Appalachians.

Unit 2 is the Carys Mills Formation containing interlayered argillaceous micrite, calcareous slate, and calcareous quartzo-feldspathic gray wacke. Locally known to geologists as "the ribbon-rock," Unit 2 is part of the Aroostook-Matapedia litho-tectonic unit. The unit extends northeastward from Presque Isle to the tip of the Gaspé Peninsula, but to the southwest, exposures die out in about 50 kilometers. These sparingly fossiliferous rocks are mostly Early Silurian (Llandoveryan) but are locally as old as Upper Ordovician (Ashgillian) and possibly older.

Rast and Stringer (1974) have suggested that Unit 2 was deposited on the floor of a small ocean basin whose eastern margin was subducted under the volcanic arc of the Tetagouche belt in New Brunswick. At the time of the authors work, the relationship between Units 1 and 2 was unknown. Although closely juxtaposed in the Presque Isle area they are everywhere separated by exposures of Unit 3. Field relationships to the north and west indicate that Unit 2 was deposited conformably on the Madawaska Lake Formation, which in turn was deposited on the rocks of Unit 1 (Roy and Mencher, 1976).

Unit 3 is a complex of Silurian clastic formations. These range in age from late Early Silurian (Late Llandoveryan) to Upper Silurian (Wenlockian, Ludlovian and locally Pridolian). D.C. Roy (see Introduction) has worked out the complex facies relationships within this Unit. He interprets it as representing a series of submarine fans derived substantially from a source-area developed on Unit 1 and deposited in a basin opening eastward. The contact between Unit 1 and Unit 3 is an angular unconformity with a hiatus spanning much of the Early Silurian. The contact between Unit 2 and Unit 3 is gradationally conformable with no discernable structural break. The stratigraphic relationships demonstrate the Units 1 and 2 were juxtaposed in their present position no later than the Late Llandoveryan. Rocks more or less comparable to those of Unit 3 are exposed for hundreds of kilometers southwest of Presque Isle as the formations comprising the western flank of the Merrimack Synclinorium.

Unit 4 comprises volcanic and sedimentary rocks of Early Devonian (Upper Gedinian) age. The volcanic rocks (Hedgehog and Edmunds Hill Formations) are distinctly more felsic than those of Unit 1. The Edmunds Hill is one of the few andesite units described in New England, although this assignment should be verified by modern chemical analyses. The volcanic rocks of the Hedgehog Formation are described as andesite, trachyte, and rhyolite (Boucot, and others, 1964). The volcanic rocks are overlain by and interfinger with the Chapman Sandstone which is also Upper Gedinian.

The interfingering is especially well-displayed at Edmunds Hill (Stop 8), where a thin tongue of Chapman Sandstone is underlain by Hedgehog Volcanics and overlain by Edmunds Hill Andesite. Fossils in the Chapman are a mixture of plants and coarse-ribbed shells suggestive of a high-energy, near-shore beach environment. At the Grindstone locality 9km south of Edmunds Hill, the Chapman contains more delicate, fine-ribbed shells of the same age as the fauna at Edmunds Hill, but suggestive of a low-energy depositional environment. There too the water also must have been shallow because a rich assemblage of plant fossils is also found. Further south, The Chapman Sandstone grades into the fine-grained Swanback Slate which may be a deeper water deposit. The relationships suggest a volcanic island in the Mapleton area, with a beach at Edmunds Hill, and a basin deepening to the south. (A visit to the Grindstone locality is highly recommended as part of a tour of the Presque Isle area, but would be too time consuming to include in the present trip. The locality is named on the topographic sheet and may be approached from the East Chapman Road).

Units 3 and 4 are separated by a disconformity (attributed to the Salinic Disturbance; Boucot, and others, 1964) with a hiatus spanning the Latest Silurian and Earliest Devonian.

Unit 4 is part of a belt of similar Early Devonian (Upper Gedinian) volcanic and sedimentary rocks extending from the west end of Chaleur Bay to the The Forks (south of Jackman) in western Maine (Boucot, and others, 1964). From Jackman to north of Mt. Katahdin this same belt is the locus of a series of caldera complexes containing thick deposits of rhyolite ignimbrite dated as Late Early Devonian (Early Emsian). These include the Traveler Rhyolite described by D.W. Rankin elsewhere in this Guidebook.

Unit 4 is thus part of a volcanic belt of substantial magnitude and considerable lateral extent. The rocks are distinctly more felsic than the earlier volcanics of Unit 1. One interpretation is that the Unit 4 volcanics evolved on a block with at least partly-developed continental affinities, whereas the Unit 1 volcanics were developed in a more oceanic environment.

Brown (1979) determined paleomagnetic pole positions for samples of the Unit 4 volcanics and noted their similarity with pole positions from volcanics of the Late Devonian Perry Formation on either side of Passamaquoddy Bay in coastal Maine and New Brunswick. Despite the significant difference in age between Early and Late Devonian she deduced in effect that the eastern and western flanks of the Merimack Synclinorium were part of a coherent block* by Devonian time. She further noted that the poles for this block differ significantly from those of Catskill and other Late Devonian units deposited on certain North-American basement. She argued from these data that the Presque Isle area was not part of North American plate in Early Devonian time.

* The author considers her designation of this block as Avalonian to be a serious misnomer. None of her samples come from the Avalon Province proper and the author has argued elsewhere that the term "Avalonian" should not be applied west of the Dover (Nfld.) and Bloody Bluff - Lake Char (SE Mass. and Ct.) Faults

The present author notes that the youngest marine Devonian rocks (including the Middle Devonian Mountain House Wharf Limestone of the Memphremagog area, southern Quebec) in the Northern Appalachians occur in the Connecticut Valley - Gaspé Synclinorium west of the Presque Isle area. It is possible that a true ocean basin persisted there in the Late Early and Early Middle Devonian. It is further possible that subduction of this block eastward beneath the rocks of Unit 1 played some role in the evolution of the volcanic rocks of Unit 4.

Unit 5 contains red, arkosic sandstone and conglomerate of the Mapleton Formation yielding plant fossils dated as late Middle Devonian (Early Givetian). These rocks (Stops 9 and 10) rest with angular unconformity on Unit 4 and contain clasts of Unit 4 and older rocks. This demonstrates that an early phase of the Acadian Orogeny occurred between Mid-Early and late-Middle Devonian time (Boucot, and others, 1964). The Presque Isle area is currently the only locality in the Northern Appalachians where a significant Acadian unconformity can be bracketed so closely.

It is noteworthy that the Mapleton Formation is itself folded, and faulted, with dips up to 50 degrees (part of which, but not all, may be original dip of the coarser sediments). Boucot and others (1964) noted that Carboniferous rocks at Plaster Rock, New Brunswick (about 80km east of Presque Isle) are flat-lying and tentatively cited. This as evidence for the presence of later phases of the Acadian Orogeny.

MISCELLANEOUS OBSERVATIONS

The work-a-day exposures in the Presque Isle area are those occurring in the ditches along farm-roads and highways. They provide the most continuous sections for describing the units and working out stratigraphic relationships. Unfortunately they come and go depending on the industriousness of the road-maintenance folk and the severity of cloudbursts. Stop 5 is an exposure of this type. It was very well exposed in 1961 when used as the type section for what now the New Sweden and Lower Jemtpland Formations. It was only moderately exposed when the road-log was prepared in August, 1979.

The potato economy of Aroostook County depends significantly on the calcareous rocks of Unit 2. Where a single farm spans the boundary between Unit 2 and Unit 3 the land underlain by Unit 2 is planted in potatoes and the remainder is used as pasture land, chicken yards, and woodlots. Potatoes can be grown on other units, but apparently with greater difficulty, as witnessed by the much greater proportion of uncleared land. The field trip route crosses this boundary on several occasions, and the participants are invited to be on the lookout for this interesting phenomenon as a relief from the tedium of driving.

Despite the extensive glacial transport of the regolith, it is interesting to note that the chemistry of the soil closely reflects the underlying bedrock. This was noted by D. Smith, who was engaged by the Maine Geological Survey in 1962, to reconnoiter the Unit 2 - Unit 3 contact from Washburn to the Canadian border. He found that the cruptic soil-chemistry codes on the Soil and Conservation Service photomosaics could be used to rough in the contact to within an accuracy of about 1/4 mile as confirmed by his subsequent bedrock mapping.

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ITINERARY

ASSEMBLY POINT - HAYSTACK MT. Maine Rt 163 (Ashland Rd.) approximately 10 miles west of Presque Isle. (Rt 163 is a left-turn off U.S. 1 about a mile north of the college.) You will see Haystack as an obvious feature on the western horizon shortly after leaving town. Park where the highway crests at the southern flank of the mountain. On the left is a gravel strip for car parking (face for departure to west); higher-slung vehicles can negotiate the road that turns right from the highway crest, crossing a small meadow to the base of the mountain.

The official trip will leave this stop at 9:00 AM. Climb as high as ambition and time permit. If you do not have time for the climb note road-cut exposures 50 m. east of the parking area, but watch traffic.

00.0 STOP 1 Felsic volcanic rocks of the Winterville Formation.

00.7 Castle Hill Picnic grounds on left

03.2 Left off Rt. 163 on dirt track leading into abandoned gravel pit. (100 yards beyond turn, Rt 163 crests a low rise, drops, then climbs steeply beyond. If you overshoot, there is a wide gravel shoulder on right for turning about 0.1 mi beyond the low crest. Drivers of low-slung cars may wish to turn here and park on south shoulder of Rt. 163. Cars will depart to east.)

STOP 2 At first sight, this appears to be a gravel pit with glacial boulders of basalt breccia, but on closer inspection one can see that the operation has scraped down to bedrock at several places near the back of the quarry. One exposure is of richly fossiliferous mafic tuff. The fossils are considered Ashgillian (Late Ordovician: R.B. Newman, oral communication, 1980).

05.8 Castle Hill Picnic ground on right.

06.5 Rt. 163 crests at Haystack Mtn.

09.6 Left off Rt 163 on paved cutoff behind Exxon Station where Rt 163 curves to right.

09.7 Left (West) on paved road

09.9 Straight (West) at jct.

10.6 Straight (West.) Dudley Farm both sides of road. Site of early 1940's test pits exploring low-grade manganese deposits in the New Sweden Fm (lower Perham) and limestone-breccia "reefs" at the base of the Jemtpland Fm (upper Perham). Note paucity of potato-fields to west in land to west underlain by non-calcareous rocks below the Jemtpland Formation.

11.7 Left turn (North) onto Turner Road near site of former Pyle School.

- 12.3 Small gravel pit on right exposes red shale of New Sweden Fm (Lower Perham).
- 14.4 Straight (North) at stop sign crossing Maine Rt. 227
- 14.5 Veer left (North) on gravel road
- 15.8 Straight (North) Summit of Richardson Hill. Site of good section through Castle Hill Volcanics and overlying units before "improvements" raised road.
- 17.8 Park on right at bottom of hill before road curves right (East). Walk (North) to base of field on left to Aroostook River (beware furrows and holes hidden by over growth); west about 0.1 mile along river to prominent outcrop.
- STOP 3 The exposure displays a gentle anticline in the sandstone and shale member of the Frenchville Formation. Note also the graded bedding. This is the most northerly exposure of the unit on the Castle Hill anticline.
- Return to cars and continue eastbound.
- 20.1 Park by low cuts on right in broad right curve in road
- STOP 4 The roadcut exposes micrite and slate of the Carys Mills Formation. Further east in outcrops along the Aroostook River (the first of which is just around the bend and down the steep bank) the percentage of micrite is higher, and the unit has the characteristic appearance to which the name "ribbon-rock" was given. The roadcut was one of the first exposures to yield fossils. They are very small graptolites, which are best seen on slabs pulled from the cut and washed by the rain. The fossils here are probably Ashgillian. The discovery of younger fossils in the river bank exposure demonstrates that the rocks locally face eastward. A fault is used to explain why the next outcrop westward is also younger (New Sweden Formation).
- 20.8 Veer left at junction with better road.
- 21.3 Straight at jct across bridge.
- 21.6 Left (Northwest) on Maine Rt. 164 into Washburn.
- 22.5 Left (West) on Maine Rt. 228 towards Perham.
- 23.8 Straight on paved secondary road as Rt. 228 curves to left.
- 26.5 Left (West) at junction.
- 27.1 Right (North) at junction.
- 27.4 Hedgerows on left (West) with field-road on north side. Make U-turn leaving room for other drivers to maneuver. Park on right facing south. Rocks for this stop exposed along field road: cars will depart to south the way they came in.

STOP 5 Boucot and others (1964) designated the section along the farm road as the type section of the Perham Formation. Roy later elevated the Perham to Group status and named the Jemmland (poorly exposed at the top of the hill) and New Sweden Formations (most of the section on the east flank). Redefinition of the gradational contact at the base of the New Sweden Formation places the lower part of the section (approximately 100 meters) in the Carys Mills Formation.

- 27.7 Right (West) at jct.
- 28.6 Veer left (West) at jct.
- 29.1 Left (southwest) on Maine Rt. 228.
Stay with Rt. 228 as it curves left then right through Perham.
- 29.4 Park on right opposite white house beyond white church at low road cut.

STOP 6 Typical siltstone of the New Sweden Formation. This is a good outcrop for working out cleavage - bedding relationships, but be cautious about what you call bedding here.

- 33.6 Stay on Rt. 228 through Washburn. Avail yourself of grocery store here if you need "fixins" for dinner (see below: in the County, "dinner" is the noon meal even if eaten out of a pail.)
- 33.8 Right (South) on Maine Rt. 164 through Washburn.
- 34.8 Right (West) off Rt. 164 across tracks and bridge.
- 34.9 Straight (West) at jct.
- 35.6 Curve left (South) with better road at jct.
- 38.5 Left (East) at jct with Maine Rt. 227.
- 38.8 Right (South) at jct taking road to Mapleton.
- 39.6 Left into Town of Mapleton Picnic Area for LUNCH STOP.
- 42.3 Left (East) on Maine Rt. 163 at jct in Mapleton.
- 43.6 Park on right at crest of hill at roadcut.

STOP 7 SPRAGUEVILLE FORMATION.

Fine calcareous siltstone characteristically showing pervasive bioturbation. (in good exposures the latter feature is useful in separating this unit from the underlying Carys Mills Formation.) The rocks are cut by teschenite (analcime-bearing, mafic) dikes that are better exposed on the hill to the south.

- 43.9 Right off Rt. 163 (South) on Pelke Rd.
- 44.1 Straight. Teschenite dikes exposed in low cuts on right.

- 44.4 Veer left (South) at jct with better road.
- 46.1 Left (East) on road towards base of Edmunds Hill.
- 46.7 Left (North) at jct. Loose blocks of fossiliferous Chapman Sandstone in rockpiles on right.
- 47.1 Park on right at gate to Trombley Quarry. Don't block entry road.
- 47.1 STOP 8 CHAPMAN SANDSTONE AND EDMUNDS HILL ANDESITE.
The best exposures are in the cleared pavement above the quarry face. In May, 1978, the interfingering of the sandstone with the volcanics was well-displayed, but the face may be expected to change for better or for worse as the quarry is worked. The two units are superficially similar (the Chapman being rich in volcanic clasts), so the eye must be attuned to the differences in texture between the igneous and sedimentary units. All of the rock types may be seen in the boulder-pile at the junction (46.7) if you cannot get into the quarry. The quarry was opened in the late 1970's.
- 47.6 Stay with main road. Low cut exposing andesite breccia. Fossiliferous pyroclastic rocks exposed in field reached by road to south.
- 48.4 Curve left (North) with main road.
- 49.5 Right (East) onto Maine Rt. 163.
- 51.1 Left (North) off Rt. 163. Park on right without blocking dump entrance. (Departure to north). Walk back to Rt. 163 and uphill (West) to small cut.
- STOP 9 Conglomeratic facies of the Mapleton Formation. PLEASE DO NOT HAMMER ON THIS CUT, and please refrain from prying clasts loose; this is the only exposure the conglomerate conveniently situated for field trips. Note the rounding of the clasts and their variety, including both sedimentary and volcanic materials. Fossiliferous clasts of Chapman sandstone have been reported. It is uncertain whether clasts of the nearby Munson's Granite have been identified; a possible candidate shown to the author by D.C. Roy about 1965 may have been a porphyritic rhyolite.
- 52.7 Curve right (North) at jct.
- 52.8 Right (East) at crossroads.
- 53.1 Park on right. Walk south into woods to small quarries.

STOP 10 MAPLETON SANDSTONE plant fossils. This locality at the crest of Winslow's Hill lies near the axis of the syncline that folds the Mapleton formation. The woods to the south of the road contain numerous small quarry-pits, and plant fossils are abundant at and near several of these. Plant micro-fossils are also abundant in these rocks.

Additional exposures of Mapleton Congl. about 1 mi. further along road.

Turn back. Retrace Route to Presque Isle Dump (see 51.1)
Left (East) onto Maine Rt. 163 into Presque Isle.

TRIP B-8

WISCONSINAN GLACIATION OF NORTHERN AROOSTOOK COUNTY, MAINE

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Introduction

The purpose of this trip is to examine multiple till exposures, stratigraphic relationships, and glacially overridden deposits which permit an interpretation of the mode of till emplacement of late Wisconsinan deglaciation in northern Maine. We will make a west to east transect (Fig. 1, St. Francis to Grand Isle) along the southern bank of the St. John River where Wisconsinan glacial deposits are readily observable.

Maps: St. Francis, Winterville, Eagle Lake, Fort Kent, Frenchville, Grand Isle, Van Buren, 15 minute series.

Stratigraphy

Surficial stratigraphic units have been mapped along the Maine side of the St. John River and yield evidence for two distinct glacial phases represented by two tills and associated outwash. The upper, or Van Buren Till, is the surface unit in northern Aroostook County. This till terminates at, and is included in, a moraine complex extending discontinuously from St. Francis to Grand Falls, N.B. (Fig. 1). Underlying the Van Buren Till at several localities along the St. John River, the St. Francis Till records an earlier event.

Lithology and Provenance

These two tills and their associated outwash deposits contain different lithologies. The Van Buren drift is characterized by inclusions of Canadian Shield Precambrian granite gneiss clasts (varying between 2 to 5%) and other exotics. The St. Francis drift is almost totally supported by the local and extensive Seboomook Formation of Devonian age (Fig. 2). No granite gneiss inclusions have been identified in the

* Work supported by the Maine Geological Survey

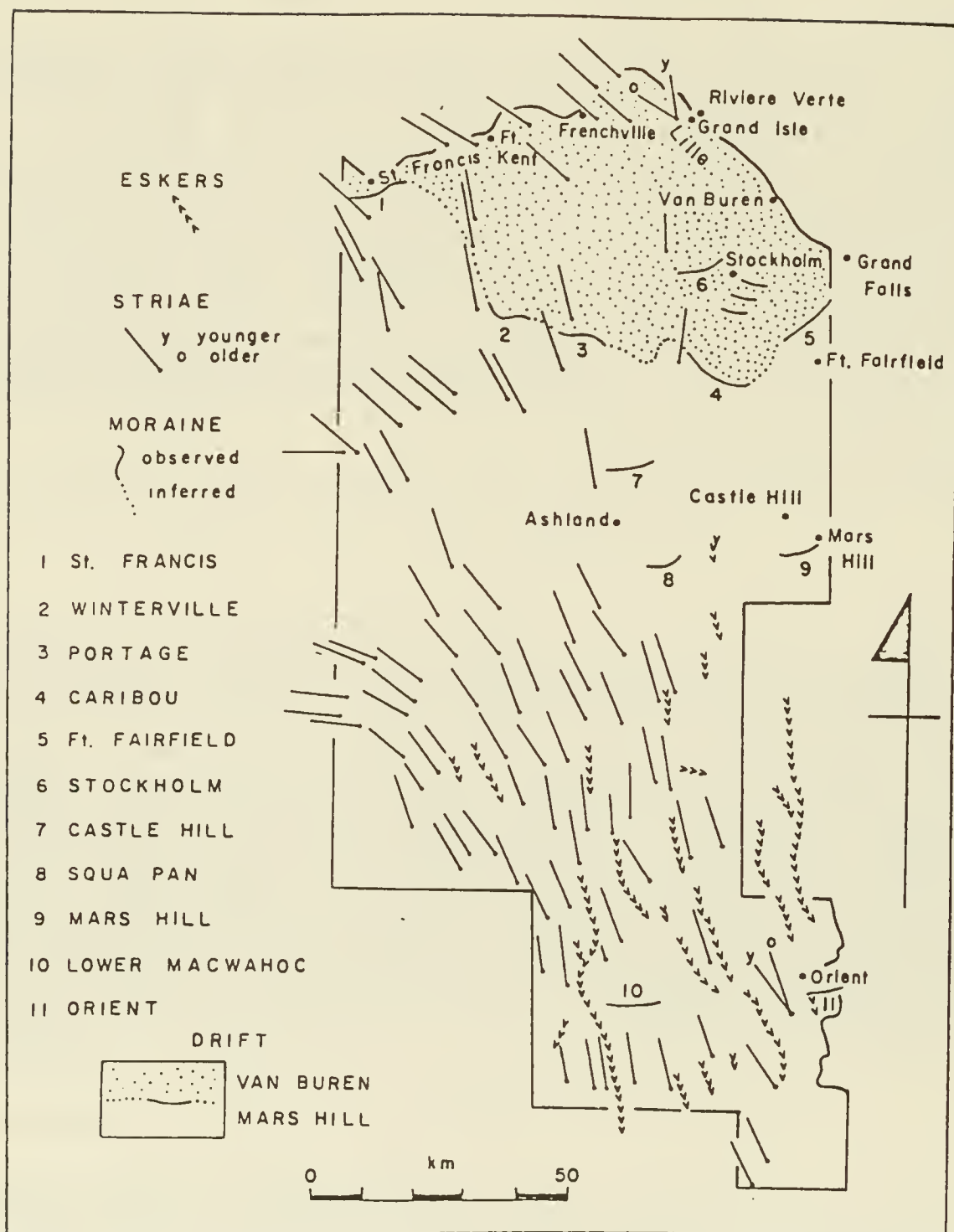


Fig. 1. Generalized surficial geology of Aroostook County.

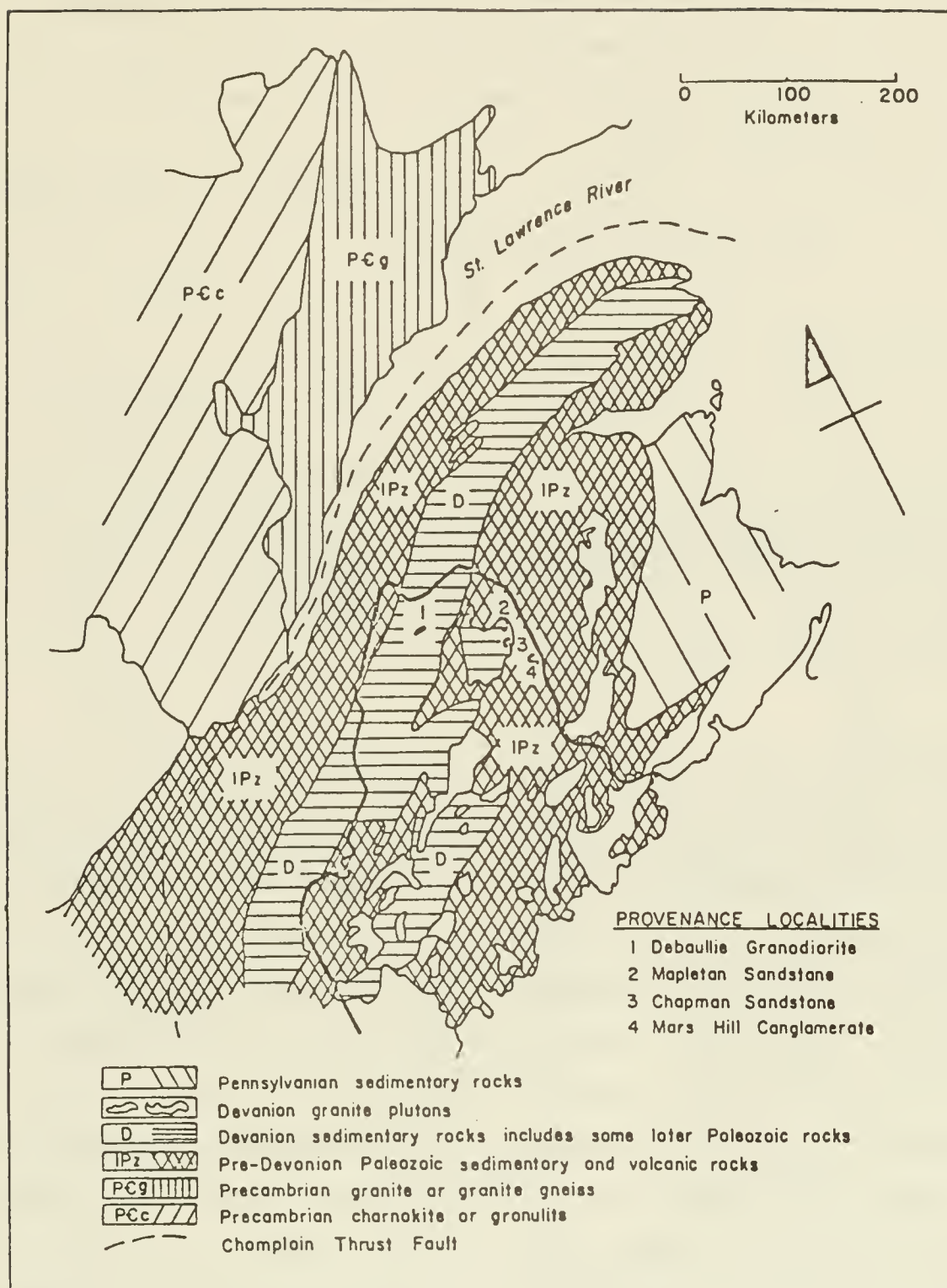


Fig. 2. Generalized geologic map of northern Maine and adjacent Canada.

St. Francis drift. Erratics, glacial striations, and landforms indicate that the Van Buren Till was emplaced by ice moving in a southeasterly direction. Till fabric of the St. Francis Till suggests lateral east-west movement. Located stratigraphically below and containing different lithologies than the Van Buren Till, the St. Francis Till is not considered to have formed from the same ice regime responsible for the deposition of the Van Buren Till.

Chronology and Correlations

We are primarily concerned with the question of how these two tills correlate with other tills identified in areas adjacent to Aroostook County and to the surface till south of the moraine complex, the Mars Hill Till. The primary stratigraphic/chronologic difficulty which is recognized results from the presence of granite gneiss clasts in the Van Buren Till and their absence in both the underlying St. Francis Till and Mars Hill Till to the south.

The difficulties in the correlation are as follows: if the Van Buren drift was deposited by ice originating north of the St. Lawrence River, the gneiss must have crossed the valley prior to the incursion of the Champlain Sea at 13,000 years BP., because there is no evidence of an ice advance through the Champlain Sea subsequent to this time. However, it also appears that the deposition of the Van Buren Till must have occurred after the formation of the coastal moraine belt at Pineo Ridge at about 12,700 years BP. because Aroostook County was almost certainly covered with ice until deglaciation followed the deposition of the coastal moraines. It would thus appear that the Van Buren Till is older than 13,000 years BP., but younger than 12,700 years BP. Even if the opening of the Champlain Sea and the deposition of the coastal moraine belt are judged to be synchronous, it is still difficult to explain the transportation of the granite gneiss to the Van Buren area by an ordinary re-advance of ice from north of the St. Lawrence Valley.

Fig. 3 shows possible correlations of Aroostook County deposits and surrounding area deposits with the stratigraphy of southeastern Quebec. The time constraints imposed on the correlations are those discussed

Stratigraphic Column, South-eastern Quebec, (McDonald and Shilts, 1971)		Stratigraphic Columns of Aroostook County, Maine, and Pertinent surrounding areas.				
Time-Stratigraphic Unit	Rock-Stratigraphic Unit	A VB MH	B VB MH	C VB MH	D MH VB	E VB=MH
Wisconsinan Stage	late	Post Lennoxville Sediments			Van Buren Grand Falls ?	
		Lennoxville Till	Van Buren Grand Falls Highland F. St. Antonin Pineo Ridge	Van Buren Grand Falls Highland F. St. Antonin Pineo Ridge	Mars Hill Highland F. St. Antonin Pineo Ridge	Mars Hill Highland F. St. Antonin Pineo Ridge
	middle	Gayhurst Formation	Outwash over St. Francis Till		Outwash over St. Francis Till	?
		Chaudière Till	St. Francis Till Mars Hill Till		St. Francis Till Van Buren Grand Falls	St. Francis Till
	early	Massawippi Formation		Outwash over St. Francis Till		Outwash over St. Francis Till
		Johnville Till		St. Francis Till		St. Francis Till

Fig. 3. Till correlation chart.

above and the date of 12,700 years BP. for the formation of the Highland Front Moraine in Quebec. (See, Introduction, Outline of the Pleistocene geology of northern Maine and adjacent Canada, this volume). Overrun sediments along the southern bank of the St. John River Valley and stratigraphic relationships indicate that the Van Buren Till is younger than the St. Francis Till at the localities where they are juxtaposed. If both the Van Buren Till and the Mars Hill Till are interpreted as being approximately the same age (Column E, Fig. 3) and both correlative to the Lennoxville Till of Quebec, then the time restriction imposed by the opening of the St. Lawrence Valley to the Champlain Sea becomes a manageable problem.

Because of markedly different sediment parameters between the two surface till units, and other considerations, the Van Buren and Mars Hill Tills are interpreted as having been deposited penecontemporaneously as the result of coalescing ice sheets (Genes and Newman, 1979) or, as the result of the thermal regime existing within a single ice mass (Hughes, pers. comm.).

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Itinerary

Mileage

- 0 Assembly point for the trip is at the IGA parking lot, Fort Kent (Junction of routes 11, 1, 161). Starting time 8: A.M. Turn west out of parking lot on to route 161.
- 0.4 American customs on right

- 0.6 St. John River flood control project on right constructed by U.S. Army Corps of Engineers.
- 1.5 Flood plain of the St. John River. Note terraces and incised outwash deposits.
- 5.3 Crossing the St. John town line.
- 6.6 On left is the farm of Sylvio Martin, seed potato farmer. No trespassing without permission.
- 13.6 Outcrops of Seboomook Formation on left. This slate underlies most of the field trip area.
- 18.6 Rankin Rapids campground on right.
- 19.9 Entering the St. Francis Moraine complex. Some bedrock is evident, however, most of the hills are not bedrock cored.
- 21.9 Chamberlain general store on left. Ernie Chamberlain was instrumental in trailblazing for the lumber companies in years past and is one of the well known residents of Aroostook County.
- 22.2 Stop 1. Park along highway. Walk down path behind cabin to the St. John River. Walk east (crossing a small stream) approximately 200 yards.

This locality (Golden Rapids) exhibits the entire section of known Wisconsin glacial stratigraphy in northern Maine. Resting on bedrock outcrop of the Seboomook Formation is the compact, silty to clayey, dark grey St. Francis Till, 3-3.5 meters thick. Clasts in this till are almost wholly derived from the underlying slate. Associated outwash, 5 meters thick, is lithologically similar to the till it overlies. No granite gneiss occurs in this till or outwash. Above the outwash is the Van Buren Till which is clayey to silty, compact, and buff to dark brown. This till and its associated outwash contains clasts of granite gneiss and other exotics. The St. Francis and Van Buren Tills, where they are juxtaposed, are separated by a barely discernible boulder pavement which consists of Seboomook Formation clasts. Near the top of the section (across the road from where the cars are parked) a third till occurs , capped by yet another thin outwash deposit. Because of this singular occurrence and similar lithologic constitution, this till is assumed to be part of the Van Buren Till (ablation facies or slight fluctuation?) and not reflective of a separate stratigraphic event. As slumping has occurred at various places at this locality, care should be exercised in examining the section.

Return to cars promptly. Reverse direction and head east.

- 22.4 Note outwash on left. This is the upper outwash of the stratigraphic section seen on stop 1. The outwash contains granite gneiss clasts.

22.6 Sand dune field on left. (Stop, time permitting).

24.1 Stop 2. Kame complex. Turn right into pit.

We are standing at the distal margin of the St. Francis Moraine complex. This large kame, some 80' high is in contact with a thick sequence of Van Buren Till immediately to the west. Because of excavation, good exposures show beds dipping outward around the entire perimeter of the deposit. The lithologies contained at this locality are those associated with Van Buren Tills and outwash deposits. Granite gneiss clasts in excess of half a meter have been found at this site.

Return to cars. Turn right out of pit and continue east.

24.3 The dirt road on the right between the two green houses (green in 1979) extends through the St. Francis Moraine complex for approximately one half mile. Pitted, hummocky, morainal topography characterizes this region. West of this hummocky moraine outwash extends to the St. John River and caps the Van Buren Till seen at stop 1.

26.3 Stop 3. Turn left into Rankin Rapids campground. Drive 0.1 mile to the large boulder at the end of the road. Go down the path to the St. John River.

Directly at the base of the path is an exposure of the St. Francis Till overlain by outwash of the Van Buren Till. Note the absence of granite gneiss or " colorful " lithology of the St. Francis Till. At times of low flow of the St. John River, it is possible to see the St. Francis Till resting directly on the Seboomook Formation (approximately 50 meters to the east). Here, till fabrics yield a strong 90° maximum.

Return to cars and return to main road (route 161).

26.5 Proceed east on route 161.

29.1 St. Francis town line.

45.1 Continue on route 161 east towards Madawaska. Fort Kent Blockhouse on left. The blockhouse marks the formal termination of canoe trips along the famed Allagash River.

45.7 Junction. Continue on route 1 toward Madawaska.

54.7 Frenchville town line.

54.9 Turn right just before the white house with the barn which is about to fall down.

55.1 Stop 4. Watch out for trucks-this is an active pit.

Excavation has destroyed much of this section, but relationships are still clearly visible. Proglacial (?) Van Buren outwash

has been overridden resulting in recumbent folding and thrust faulting on a large scale. Photographs of the original section will be available if the pit is virtually destroyed by the time we get there. This locality is directly north of the Caribou Moraine (to be visited on trip C - 6) and is thought to represent the event responsible for the formation of the moraine.

Return to cars and return to the main road.

- 55.4 Proceed south on route 1.
- 58.6 Turn right onto road to northern Aroostook Airport.
- 58.7 Turn right at sign to airport. Continue on main road.
- 62.4 Take left fork by stop sign. Long Lake is on the right. Continue on main road.
- 64.9 Take right on Grand Isle road just before the EXXON sign in front of the general store.
- 71.9 Take right at four corners. White house on hill to the right. Across from the white house is a two striation locality (possible stop).
- 75.2 Turn left at main road. Entering Lille sign to the right.
- 75.4 Stop 5. Turn left into gravel pit. Light green house just before entering the pit.

Here, the Van Buren Till, 3-5 meters thick, has overridden outwash deposits associated with the St. Francis Till. The outwash is rasped, torn, and large inclusions of the underlying deposit is included within the overlying Van Buren Till.

Return to cars and head for home.

End of Trip

DEGLACIATION OF THE EDMUNDSTON AREA
AND REAPPRAISAL OF GLACIAL LAKE MADAWASKA INTERPRETATION

Claude Gauthier (Terrain Sciences, Geological Survey of Canada) and Jacques Thibault (Dept. of Natural Resources, New Brunswick)

The history of Glacial Lake Madawaska is closely linked to the last deglaciation in Central and Northern New Brunswick as well as in Northern Maine. New evidence suggest its outlet may have been to the north, across the Appalachian Mountains, into the St. Lawrence Valley (Gauthier, 1980). This hypothesis will be investigated during the coming 1980 field season and in the light of mapping by Thibault (1979, 1980). The main implications of a northward outlet are: 1) persistence of ice in Central New Brunswick and the consequently limited (if not inexistent) influence of the Laurentide ice during deglaciation, and 2) damming of the St. John River by ice in the Grand Falls region and resulting in the the formation of Glacial Lake Madawaska.

Chalmers (1885, p. 41-42 GG) described "chains of lakes" formed in Saint John Valley behind drift dams (the most prominent one being a frontal moraine (Lee, 1959) at Grand Falls). Breaching of the drift in the valley gradually lowered the lake level and drainage was established along a new course. Kiewiet de Jonge (1951) traced the extent of this lake up to Lac Témiscouata, Quebec and gave it the name: "Glacial Lake Madawaska"; he recognized only one single phase related to the existence of the lake. Lee (1953, 1955) confirmed this interpretation and traced the extent of the shoreline deposits in the region of Edmundston. Martineau (1979) observed rhythmite deposits in the Lac Témiscouata area; he also reported the existence of two opposite glacial flows in the area: an early flow towards the southeast (Laurentide ice) followed by a flow with reversed direction (Appalachian ice). Martineau did not present the potential regional implications of his finding.

In August 1979, during a one day field trip excursion in the area around Edmundston, several striation sites were observed. Convincing evidence of striated and polished outcrops indicate that glacial flow was active towards the west and the northwest, controled by the orientation of the Saint John Valley and some of its tributaries. Sites were observed along the Madawaska and the Iroquois valleys (six sites with northwest flow) as well as along the Saint John River near the city of Edmundston (three sites indicated a westward flow). No other flow direction was noted, and all striated outcrops reflected a single ice flow direction (with maximum variations in the order of 20 degrees). Sense

of flow was obtained from numerous and distinctive characters: on a microscale, crag-and-tail and nailhead features, and plucking of lee sides of outcrops; on a macroscale, well developed stoss-and-lee relationships on outcrops, producing whalebacks and roches moutonnées. Although the number of observation sites is small, it is believed on the basis of 1) the uniformity of the various observations, 2) the intensity of the erosional features, and 3) the absence of other movements, that the stiae represent the last glacial flow in the area, and that this flow was channelled by the major valleys of the area.

Implications of the New Observation

The present observation suggests a new interpretation of the mode of deglaciation of the Saint John Valley and of the origin of Glacial Lake Madawaska. In order to generate a northwestward-moving ice mass, the glacier would have had to be centred over the highlands of central New Brunswick (and Maine?) and it would have had to flow radially in several directions. At one time, the frontal position of the ice was located in the Grand Falls area, forming the so-called Glacial Lake Madawaska by ice damming the valley. Subsantiating evidence has been collected on the east side of the hypothetical icemass in the Bathurst area (Gauthier, 1979 and 1978).

The region of Plaster Rock (28 km east-southeast of Grand Falls) presents a set of discontinuous morainic ridges with a relief locally greater than 60 m. From limited evidence gathered in the field (one transversal section) glaciofluvial stratification in the moraines (beds dipping at 5-20 degrees towards the west) indicates ice localized to the east side of the moraines. The ice contact ridges, in spite of their high relief (greather than 60 m), have no obvious regional extent; nevertheless, they seem to represent a position of an ice cap centred in the New Brunswick Highlands.

Field trip

The trip logs will be presented at the end of the summer, following the summer's field work in the area. Conclusions will be reached at that time concerning the significance of the westward glacial flow and its regional extension. Classical sections along the St. John River at Grand Falls, striation sites, the Plaster rock moraines and related observations will be included in the trip.

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TRIP B-10

THE GEOLOGY AND DEFORMATION HISTORY OF THE SOUTHERN
PART OF THE MATAPEDIA ZONE AND ITS RELATIONSHIP
TO THE MIRAMICHI ZONE AND CANTERBURY BASIN

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INTRODUCTION

Western New Brunswick is underlain by parts of two tectonostratigraphic zones: the Matapedia Zone and the Miramichi Zone. Part of the Miramichi Zone is downfaulted forming the Canterbury Basin in the southern part of the area (Figure 1). The purpose of this trip is to examine sections from Florenceville to Canterbury and contrast the structural development of the two zones with the Canterbury Basin.

The Matapedia Zone has been mapped in the Woodstock area by Anderson (1968), and Hamilton-Smith (1972). The regional geology in the southern part of the Matapedia Zone has been demonstrated by Pavlides (1968). Reconnaissance work by us has modified the structural interpretation for this area and an early (Pre-Acadian) period of recumbent folding is recognized. Recent mapping by the Department of Natural Resources has modified previous interpretations of the stratigraphy, structure and metamorphism of the Miramichi Zone. The rocks in the Canterbury Basin have been mapped by Venugopal (1978, 1979) and Lutes (1979) and the general stratigraphy and structure are well known.

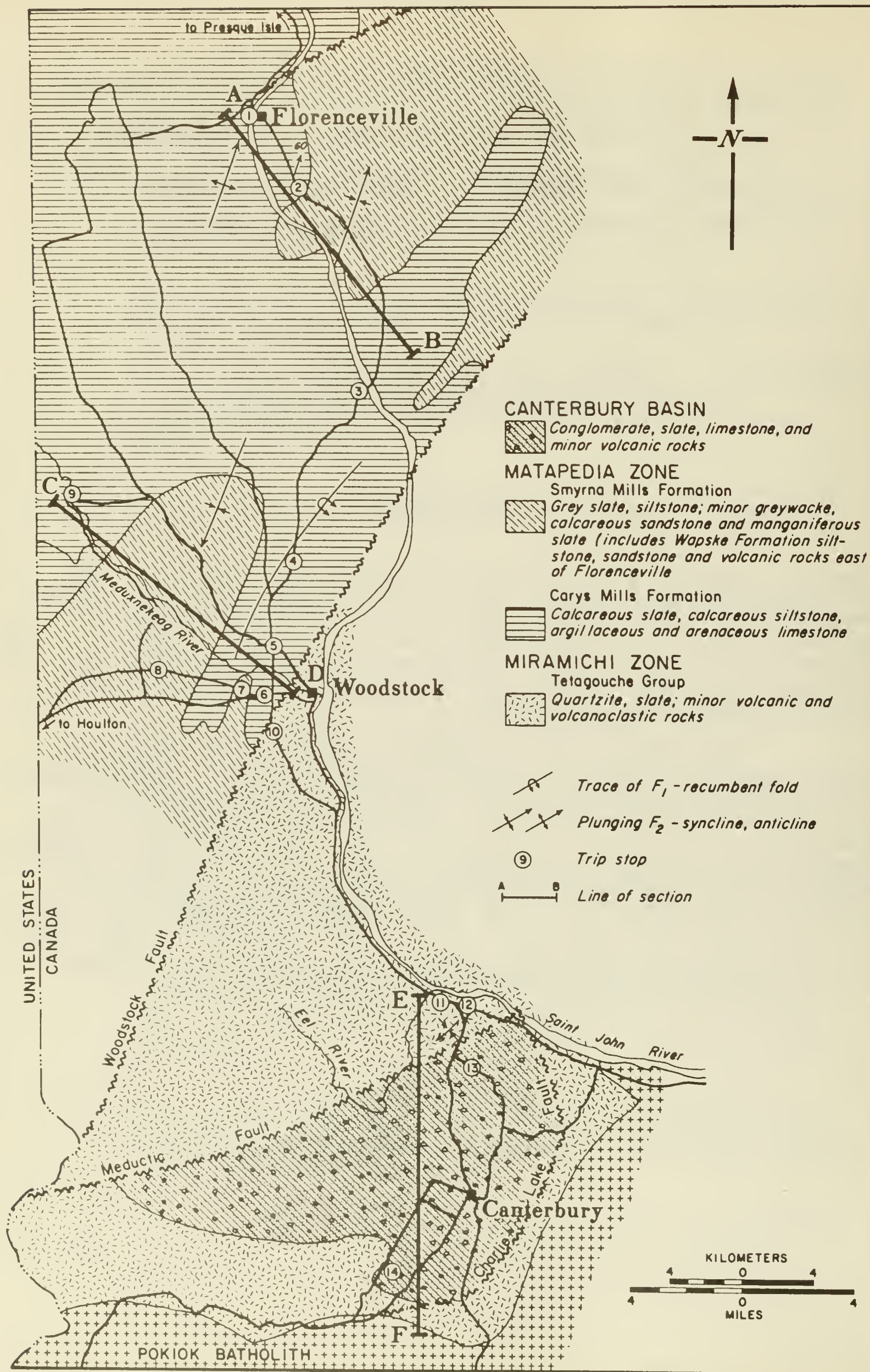


Figure 1. Generalized Geology of Western New Brunswick.

MATAPEDIA ZONE

Stratigraphy

The Matapedia Zone strikes from northeastern Maine across northwestern New Brunswick into the southern part of the Gaspé Peninsula. The zone is composed of limestones, greywackes, siltstones and shales ranging from Middle Ordovician to Late Silurian in age. The rocks in northern New Brunswick have been separated into four rock-stratigraphic units: Grog Brook Group, Matapedia Group, Upsalquitch Formation, and Perham Formation (St. Peter, 1978). In the Woodstock area, the Matapedia Zone is underlain by calcareous slates of the Carys Mills Formation which are stratigraphic equivalents of the Matapedia Group and range in age from Middle Ordovician to Lower Silurian (Pavlides, 1968). Its thickness has been estimated to range from 1500 to 12,000 feet in Maine; the formation thins to the west (Pavlides, 1968). In New Brunswick, it has been estimated to be greater than 2500 feet thick (Hamilton-Smith, 1972) and has been interpreted as a turbiditic sequence of calcareous flysch (Pavlides, 1968). The Smyrna Mills Formation conformably overlies the Carys Mills Formation in Maine (Pavlides, 1968). It is composed of slate, calcareous slate and sandstone with minor manganeseiferous siltstone. This formation ranges in age from early Llandovery to early Ludlow (Pavlides and Berry, 1966). Parts of the Wapske Formation occur above the Smyrna Mills Formation southeast of Florenceville. The Wapske Formation consists of intercalated clastic sedimentary and mafic volcanic rocks which are interpreted by St. Peter (1978) as an alternating sequence of marine and terrestrial rocks, the marine rocks probably being shelf deposits. The age of the Wapske Formation is established as Lower Devonian, probably Helderbergian (St. Peter, 1978).

Structure

Two periods of folding have been recognized in the Woodstock-Florenceville area. The first has produced large recumbent nappes which are interpreted from flat-lying overturned beds near Woodstock. There is no cleavage associated with these F₁-folds and the age of deformation is difficult to assess. The Smyrna Mills Formation has been demonstrated to conformably overlie the Carys Mills Formation near Houlton, Maine (Pavlides, 1968). Near Woodstock (Stop 7), the contact between the two formations appears to be conformable, suggesting at least part of the Smyrna Mills Formation was involved in the recumbent folding. As noted previously, the youngest age known for the Smyrna Mills Formation is Ludlow (Pavlides and Berry, 1966). The oldest age for the

overlying Wapske Formation is Lower Devonian (Helderbergian). The intervening and unrecorded stratigraphic interval may represent a period during which uplift in the adjacent Miramichi Zone created unstable slope conditions in part of the Matapedia Basin initiating gravity slides prior to the main Acadian Orogeny. It is also possible that the Smyrna Mills Formation, where observed, is paraconformable over the Carys Mills Formation. If so, gravity sliding may have been initiated by Taconian movements in the Miramichi Zone in Upper Ordovician time.

A superimposed Acadian generation of folding and cleavage affects all rocks in the Matapedia Zone. The style and attitude of these folds varies with change in orientation of the earlier F_1 -folds. On the limbs of F_1 -folds, Acadian folds are open to close and have associated axial planar cleavage. Plunge of these folds is dependent on the attitude of the F_1 -fold limb. In the vicinity of F_1 -fold hinges, Acadian folds are tight and have large variations in plunge and bedding-cleavage intersection lineation.

MIRAMICHI ZONE

Stratigraphy

Rocks of the Tetagouche Group underlie most of the Miramichi Zone. The rocks range from Cambrian(?) to Middle Ordovician and can be divided into five map-units consisting of a thick basal unit of quartzite and slate(1) overlain in succession by (2)slate, siltstone and greywacke, (3)rhyolitic volcanics, (4)manganiferous slate, chert and andesitic volcanics, and (5) massive basic volcanics (Helmstaedt, 1971). In southwestern New Brunswick, lithostratigraphic equivalents of the first four units have been recognized (Venugopal, 1979; Lutes, Unpubl. Msc.). Manganiferous slate forms a readily recognizable marker horizon throughout the Miramichi Zone.

Structure

Rocks of the Miramichi Zone are structurally more complex than those of the Matapedia Zone or Canterbury Basin. First folds, which are rarely observed, have associated cleavage parallel to bedding. Later, second folds of bedding and the first cleavage in the Canterbury area pre-date the main third deformation which folds and cleaves both Cambro-Ordovician and Siluro-

Devonian rocks (Lutes, 1979). These third folds are generally steeply dipping with axial planar cleavage. Much of the Miramichi Zone is occupied by granitoid plutons, generally of Devonian age. Whereas regional metamorphism has little affected rocks of the Matapedia Zone and the Canterbury Basin, the grade of metamorphism in the Miramichi Zone reaches sillimanite grade within a megmatite complex that occurs in central New Brunswick. Deformation and regional metamorphism of this zone is attributed to the Taconian Orogeny of Middle-Upper Ordovician age and spans the time during which the Carys Mills Formation was being deposited in the Matapedia Basin.

CANTERBURY BASIN

Stratigraphy

Rocks of the Canterbury Basin comprise conglomerates, slates, mafic and felsic volcanics and limestone unconformably overlying rocks of the Miramichi Zone. These Siluro-Devonian rocks occupy a downfaulted basin bound by the Meductic Fault on the west and the Charlie Lake Fault on the east (Figure 1). Pocowogamis Conglomerate forms the base of the succession along the east side of the Meductic Fault. This is overlain by Scott Siding Slate succeeded by feldspar-quartz crystal tuff and mafic volcanics. In the southern part of the area, Cambro-Ordovician quartzite of the Miramichi Zone is unconformably overlain by Canterbury Limestone with some basal conglomerate. The Canterbury Limestone and Scott Siding Slate are gradationally overlain by calcareous sandstones and siltstones of the Hartin Formation containing intercalated mafic and felsic volcanics, microconglomerate and minor limestone. Fossils from the Hartin Formation have yielded a Lower Devonian (Helderbergian) age. Underlying formations may be Silurian.

Structure

Rocks in the Canterbury Basin occupy a downfaulted basin (graben structure) and subsequent deformation produced north facing folds in the south and south facing folds in the north. First schistosity in these rocks has a similar trend and attitude as second schistosity in Cambro-Ordovician rocks and is probably the same age. Metamorphism by the nearby Pokiok Batholith has formed biotite in many of the Siluro-Devonian rocks on the north.

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ITINERARY

Assembly point is in Florenceville at stop 1, a deep road cut on route 103 about 0.5 km southwest of the intersection with route 2 on the west side of the bridge to East Florenceville across the Saint John River. Park vehicles along side of road. Assembly time is 9:00 a.m.

Kilometers	Miles	
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00.0	00.0	<u>Stop 1.</u> Carys Mills Formation. The beds are closely folded (of Acadian generation) and plunge approximately 55° to 016° . Grading and small-scale load structures suggest bedding is right way up and faces northward. Numerous faults cut the section and gabbroic dykes are common.
00.5	00.3	Intersection with rts. 2 at bridge. Proceed across bridge and travel southward on rte. 2.
06.1	02.4	<u>Stop 2.</u> Small exposure on east side of road. Carys Mills Fm. Thinly bedded and laminated calcareous siltstones and slate. Grading appears to be right way up on cleavage trending 014° and dipping vertically. Bedding-cleavage intersection plunges about 60° to 014° . Structural style appears to be consistent to this point. Over the next 6-7 km we will be crossing a section of Smyrna Mills Fm. and Wapske Fm. which overlie the Carys Mills Fm. and appear to plunge northwesterly.
20.1	12.6	<u>Stop 3.</u> Carys Mills Fm. Current laminated, thinly bedded calcareous sandstone and lithographic limestone. Bedding consistently trends 020° , dips 50° to the west and is right way up. Bedding-cleavage intersection is subhorizontal and indicates a change in the plunge of Acadian folds. As the Acadian folds are superimposed on an earlier (F_1) generation of recumbent folds, this must reflect a change in the attitude of first structure. We are interpreted to be on the western limb of an overturned recumbent F_1 -fold (Figure 2a,b).

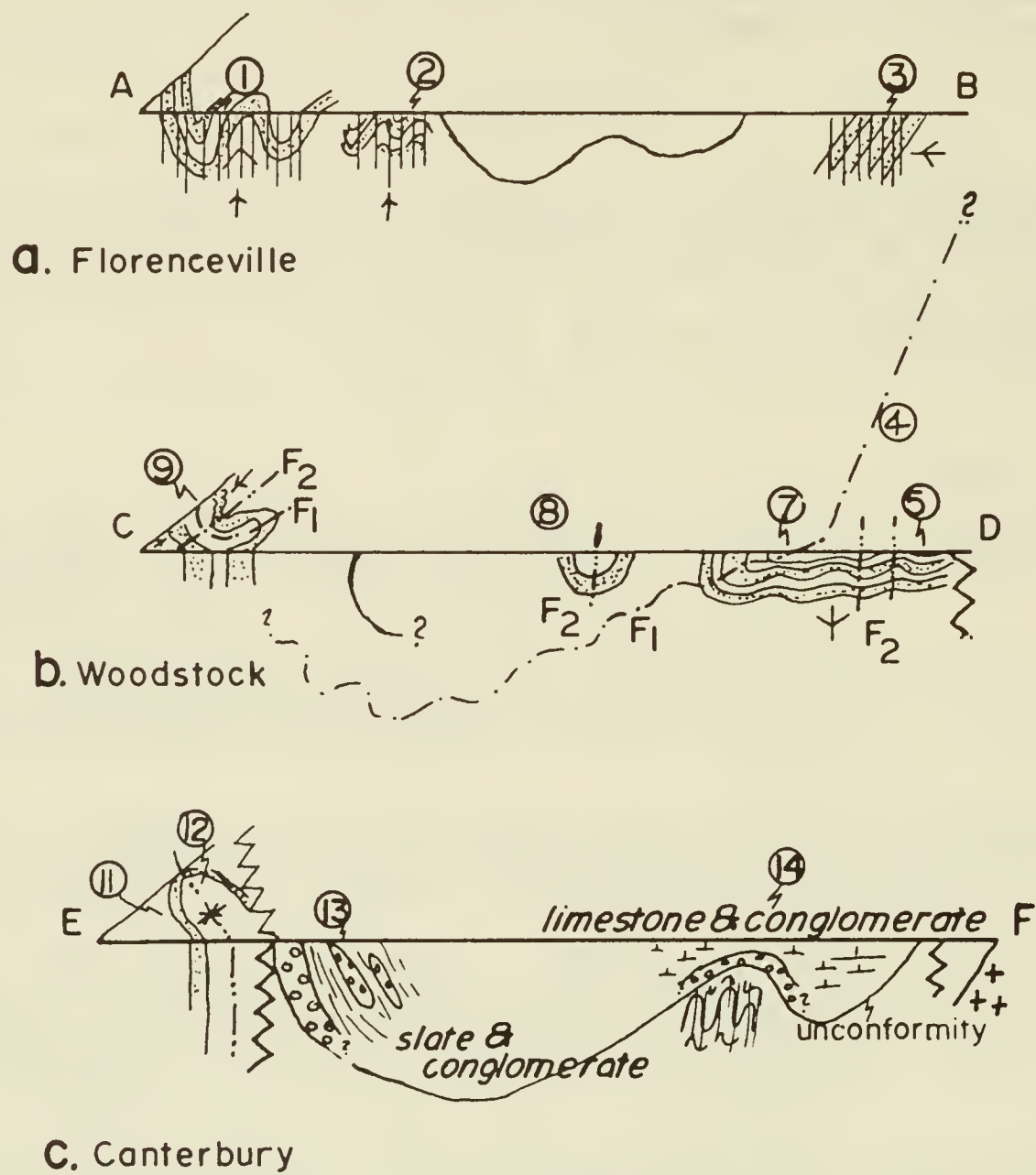


FIGURE 2.

Schematic cross-sections through parts of western New Brunswick. Trip stops are indicated.

Kilometers Miles

- | | | |
|------|------|--|
| 31.5 | 19.7 | <u>Stop 4.</u> Carys Mills Fm. Calcareous slate and siltstone. Small isoclinal F_2 (Acadian) folds with axial planar cleavage trending parallel to road (024). The plunge of these folds is extremely variable within the outcrop as is indicated by bedding-cleavage intersections. This appears to be the affect of superimposed folding over F_1 -folds. We are interpreted to be in the hinge area of a major F_1 -fold (Figure 2a,b). Younging indicators are inconclusive, but exposure 1 km to the south appears to be downward facing. |
| 32.6 | 20.4 | F_2 -folds in this exposure plunge about 20° to 204 and bedding appears to be downward facing. |
| 38.5 | 24.1 | <u>Stop 5.</u> Intersection of rte. 2 and hwy. 550. Carys Mills Fm. Calcareous slate and siltstone. Bedding is flat lying, downward facing and openly folded by upright, horizontal F_2 -folds (Figure 2b). Grading and dewatering structures give facing direction. We are interpreted to be on the overturned limb of the F_1 -recumbent fold. |
| 41.3 | 25.8 | Intersection of rte. 2 and rte. 95 to Houlton. Turn right. |
| 41.5 | 25.9 | <u>Stop 6.</u> Carys Mills Fm. Similar lithology and structure as stop 5. This exposure contains the best criteria for downward facing. Grading is well developed and a sedimentary cut-off provides convincing evidence of way up. |
| 42.3 | 26.4 | Intersection, bear right. |
| 43.6 | 17.0 | <u>Stop 7.</u> Carys Mills Fm. is steeply dipping, tightly folded and faces southward. An apparently conformable contact with Smyrna Mills Fm. is exposed to the west, on the south side of the highway. Smyrna Mills Fm. here consists of steeply dipping thinly bedded sandstone and slate which also appear to be tightly folded by steeply plunging F_2 -folds which face to the southwest. We are interpreted to be in the hinge area of the F_1 -recumbent fold. |

Kilometers	Miles	
48.2	30.1	<u>Stop 8.</u> Smyrna Mills Fm. Fine-grained green calcarenite and slate. Large upright syncline has very shallow plunge. Graptolites can be found in this exposure.
49.2	30.6	Intersection with rte. 540. Turn right and follow road past Belleville and Jackson Falls to Oakville.
61.7	38.6	<u>Stop 9.</u> Oakville. Turn right, cross bridge and park on east side of river. Polydeformed Carys Mills consists of argillaceous limestone with thinly bedded and laminated sandy limestone. Steep F_1 -folds are refolded by superimposed F_2 -folds and cleavage trending 010 (Figure 2b).
82.1	51.3	Turn and proceed back to Intersection of rte. 2 and rte. 95 at Woodstock, past stop 6. Turn right and travel southward.
84.3	52.7	<u>Stop 10</u> (optional). Strongly sheared quartzite, quartz wacke and slate of the Miramichi Zone southeast of the Woodstock Fault. This fault separates the Matapedia and Miramichi Zones and has a pronounced topographic expression trending in a northeasterly direction.
104.1	65.1	<u>Stop 11.</u> Polydeformed quartzite and slate just east of Meductic across Eel River. This is the typical lithology of the basal Tetagouche Gp. of northern New Brunswick and is similar to Grand Pitch type lithology in Maine. Small-scale, vertically plunging, asymmetric F_2 -folds with axial planar fracture cleavage fold bedding, first cleavage (which is sub-parallel to bedding), and L_1 -intersection lineation which, if axial planar, suggests F_1 -folds were originally plunging southwards.
105.1	65.7	<u>Stop 12.</u> Long road-cut of steeply dipping reddish, manganiiferous slate, chert and volcanoclastic sedimentary rocks. These overlie the quartzite-slate unit and are interpreted to occupy the axial zone of a steeply plunging, southward facing syncline

Kilometers Miles

(Figure 2c). More quartzite outcrops to the east, and the northeasterly trending Meductic Fault truncates the Cambro-Ordovician sequence approximately 2 km to the east and to the south.

105.2	65.7	Turn vehicles and proceed southward on rte. 122 towards Canterbury.
108.4	67.8	Intersection. Turn left toward Johnson Sett.
109.4	68.4	<u>Stop 13.</u> Small outcrop on left of strongly cleaved polymict conglomerate. Conglomerate is extensive in the Siluro-Devonian sequence of the Canterbury Basin along the Meductic Fault. Venugopal (1979) has envisaged formation of the conglomerates along an escarpment formed by the Meductic Fault. Grey Scott Siding Slate, a dominant lithology in this area, becomes calcareous to the south and grades into the Canterbury Limestone.
110.4	69.0	Return to main road, turn left and proceed to Canterbury.
119.4	74.6	Canterbury. Bear right past school and turn right past cemetery.
122.4	76.5	Intersection. Turn left and proceed southward across railway tracks.
128.4	80.3	<u>Stop 14.</u> Pass church on right side of road and park in drive of deserted house at bottom of hill on left. Exposure is scattered in back field. The rock is essentially a quartzite-pebble conglomerate with a limey matrix. This represents the base of the Canterbury Limestone and unconformably overlies Cambro-Ordovician quartzite and slate immediately to the west from which the pebbles were derived. The relatively higher topography marks areas underlain by quartzitic rocks.

END OF TRIP

Return to cars and retrace to route 2.

TRIP C-1

STRATIGRAPHY AND SEDIMENTOLOGY OF THE
SIEGAS FORMATION (EARLY LLANDOVERY) OF
NORTHWESTERN NEW BRUNSWICK

Terence Hamilton-Smith
Sohio Petroleum Company, San Francisco

Introduction

The Siegas area extends across the international border between the United States and Canada at the intersection of the Grand Isle, Stockholm and Van Buren 7 1/2 minute quadrangle maps of the U.S. Geological Survey. Geological work in the area was done between 1967 and 1969 as a Master of Science thesis (Hamilton-Smith, 1969) at the Massachusetts Institute of Technology. Acknowledgements are due to R. R. Shrock, E. Mencher, A. T. Boucot, W. B. N. Berry, J. M. Berdan, R. B. Neuman, J. W. Huddle, D. C. Roy, T. B. Griswold and G. Planansky for assistance, encouragement and guidance during the work. Financial assistance was received from the National Science Foundation and Department of Natural Resources of the Province of New Brunswick.

Rocks of the Siegas area are entirely sedimentary and are between Middle Ordovician and Late Silurian in age. They occur in the tightly folded common limb of the Ashland Synclinorium and the Pennington Mtn. Anticlinorium. The earliest tectonic event recorded in the rocks was the Taconic orogeny, which resulted in the deposition of the Siegas Formation and possible low intensity folding in some older rocks. Bulk deformation of the area occurred after the Late Silurian, probably during the Acadian orogeny.

A simplified geological map of the Siegas area is presented in Figure 1. A homogeneous fold system is recognized which consists of plane, cylindrical, tightly appressed similar folds which are slightly inclined and plunging. Details of outcrop distribution, structural attitudes, etc. may be found in Hamilton-Smith (1970).

Stratigraphy

The oldest unit exposed in the Siegas area is the Madawaska Lake Formation (Roy et al., 1976) of Late Middle to Late Ordovician in age. The unit is more than 1950 feet thick and is composed of dark grey slate with minor quartzose sandstone. The base of this unit is not exposed. The sandstone is light grey, highly calcareous (25%), quartzose (70%) and fine-grained, with minor amounts of plagioclase, biotite, lithic fragments and sphene. This lithotype occurs in beds from 1/2 inch to 4 feet thick and is usually laminated and cross-laminated. Graptolites collected from one locality in the Siegas area indicate a probable Zone 13 of the Caradoc age (Hamilton-Smith, 1970).

The Carys Mills Formation is between Late or Late Middle Ordovician and earliest Silurian in age and conformably overlies the Madawaska Lake Formation in the Siegas area. The Carys Mills Formation is about 1300 feet thick and is composed of interbedded slate, sandstone and limestone. The characteristic lithotype is dark grey, argillaceous calcilutite (75-90% calcite) which weathers either light bluish grey or light brown. This occurs in beds from 1/2 inch to 16 inches thick and is commonly crosslaminated as modified by convolute lamination. Graded bedding is locally common but obscurely developed, particularly in the finer grained beds. Conodonts and graptolites collected from two localities in the Siegas area indicate a Late Ordovician to Early Silurian (early or middle Llandovery) age (Hamilton-Smith, 1970).

The Siegas Formation of Early Silurian (early Llandovery) age overlies the Carys Mills Formation in the Siegas area with general conformity. Local unconformity due to submarine erosion is established at two localities. The thickness of the Siegas Formation ranges from 800 to 350 feet in the area. The unit is composed of sandstone with minor conglomerate, slate and limestone. Details of facies and lithotypes are discussed below. The sandstone occurs in beds from 1 to 314 inches thick with a variety of sedimentary structures suggestive of deposition from turbidity currents. Modal compositions range from quartz arenites through arkose to lithic wacke. The conglomerate occurs in beds from 2 inches to 27.5 feet thick and consists of limestone clasts up to 36 inches in maximum dimension contained in a lithic wacke matrix. An extensive collection of brachiopods from the Siegas Quarry (EM558) indicates an Early Llandovery age (Ayrton et al., 1969).

The New Sweden Formation (Roy et al., 1976) conformably overlies the Siegas Formation and is between Early Llandovery and Ludlow in age. No fossils are known from this unit in the Siegas area. The New Sweden Formation is about 600 feet thick and consist of grey calcareous slate with minor mamganiferous siltstone.

The Jemtland Formation (Roy et al., 1976) conformably overlies the New Sweden Formation. The unit may be from late Wenlock through early Ludlow in age but in the St. John River valley only Early Ludlow fossil collections are known. No fossils are known from this unit in the Siegas area. The Jemtland Formation is composed of calcareous shale, slate and siltstone with minor sandstone and limestone and is probably more than 2000 feet thick. The top of this unit is not exposed.

Siegas Formation

The Siegas Formation is of restricted areal extent. To the southeast the unit disappears within 8 miles, probably passing laterally into equivalent beds of the Carys Mills Formation of Early Llandovery age. To the southwest the unit may pass laterally into equivalent beds of the basal Frenchville Formation. Reconnaissance mapping suggests that the Siegas Formation extends at least 7 miles to the north and northwest and possibly as much as 40 miles to the northeast.

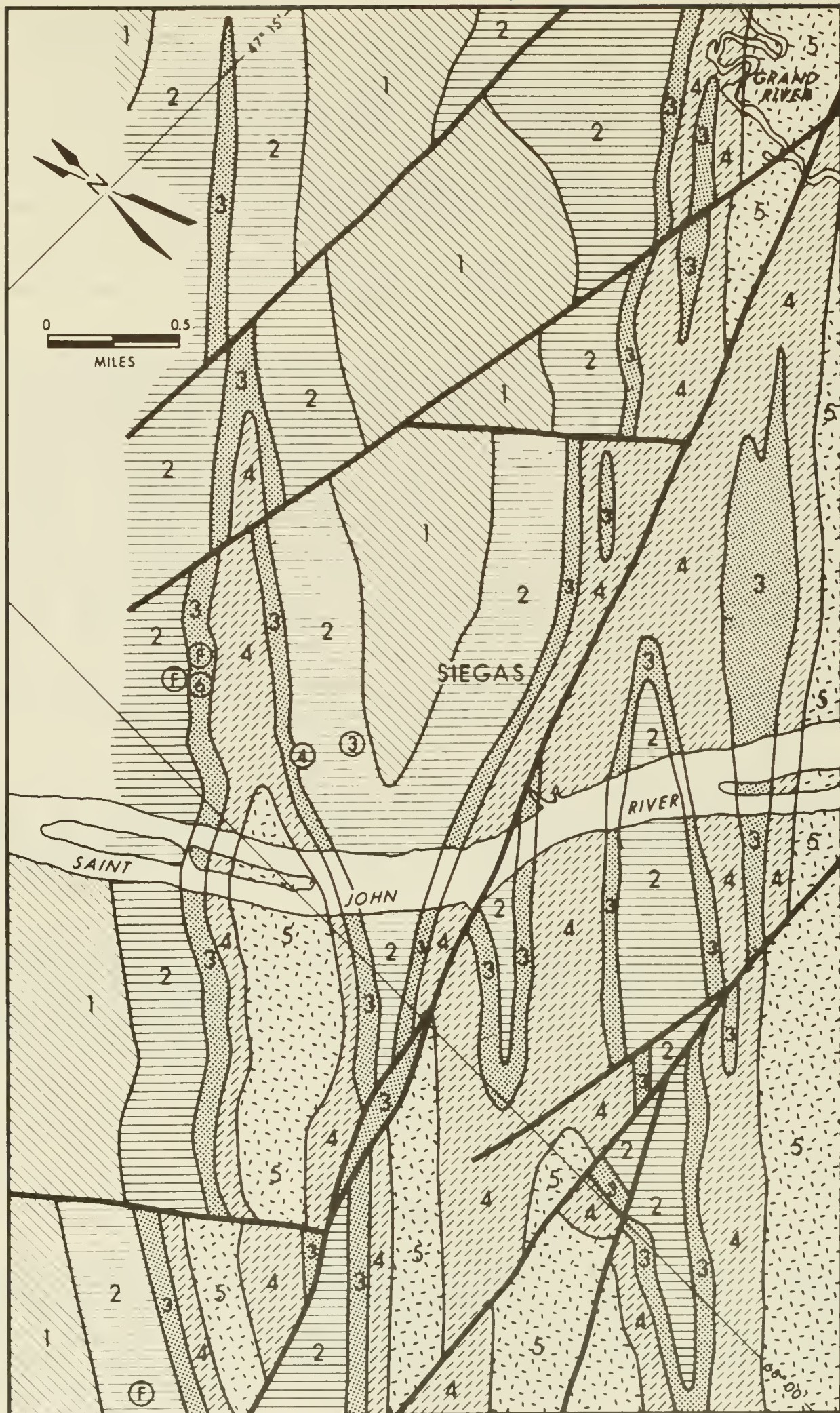
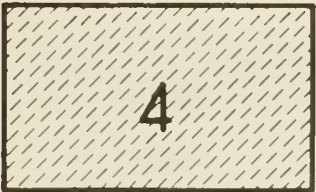


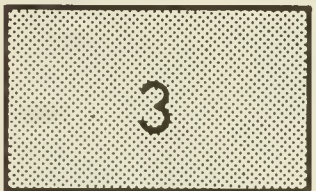
Figure 1: Generalized geological map of the Siegas area,
New Brunswick and Maine.



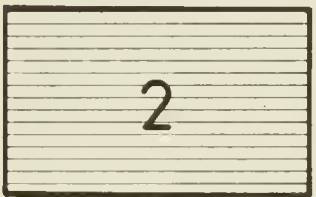
Jemtland Formation: thinly interbedded, fissile, laminated shale, slate and siltstone with minor fine grained light grey sandstone and limestone.



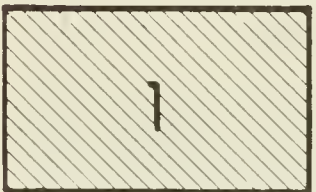
New Sweden Formation: dark to medium grey, laminated, calcareous slate with laminated manganiferous iron-rich siltstone.



Siegas Formation: interbedded sandstone, slate and conglomerate with minor limestone and chert. Sandstone composition varies from lithic wacke to quartz arenite.



Carys Mills Formation: thinly interbedded dark grey, calcareous slate and limestone. Limestone is dense and micritic and weathers characteristic blue grey and light brown colors.



Madawaska Lake Formation: dark grey, sparsely laminated, non-calcareous slate with minor light grey calcareous fine grained sandstone.



Formation contact
Fossil locality



Fault
Scheduled stop

Within the Siegas area there are significant facies variations within the Siegas Formation. The unit in the western part of the area is 600 to 800 feet thick, characterized mainly by lithic wacke and limestone conglomerate. To the southeast the unit consists of 500 to 600 feet of section, mainly sandstones with a composition ranging from feldspathic arenite to arkosic wacke. To the northeast the Siegas Formation is 350 to 500 feet thick and is composed mainly of quartz arenite, slate and limestone.

The lithic wacke facies displays significant changes from south to north. There is an increase in total section thickness from 500 to 800 feet, an increase in the thickness and abundance of limestone conglomerate beds and the progressive development of an erosional surface between the Siegas Formation and the Carys Mills Formation. As much as 150 feet of the Carys Mills Formation has been eroded in the north part of the facies, at the Siegas Quarry section. The well exposed section in the Siegas Quarry is discussed in detail below.

Facies variations within the Siegas Formation are interpreted as the result of deposition by different processes in a region of complex topography. The lithic wacke facies is understood as the product of a submarine channel-fan system. Paleocurrent information from the Siegas Quarry indicates a derivation from the north. The quartz arenite facies is interpreted as a winnowed shelf deposit resulting mainly from wave action. The arkosic facies is a finer grained deeper water equivalent of the quartz arenite facies deposited on the slope between the shelf and the submarine fan (Hamilton-Smith, 1971a).

The composition of sandstones of the Siegas Formation indicates a specific provenance. Sandstones of the lithic wacke facies consist mainly of andesite fragments and their disintegration products: plagioclase (An_{32} average), quartz and pyroxene. Pyroxene and plagioclase compositions of the sandstones are identical to those of the phenocrysts of the andesite fragments. A distinct minor petrological assemblage (up to 20%) consists of felsic plutonic fragments and their disintegration products: potassium feldspar and quartz. Quartz grains from these two assemblages are morphologically distinct.

Sandstones of the arkosic facies consist mainly of felsic plutonic fragments, potassium feldspar and quartz. The feldspars of the plutonic fragments include orthoclase, perthite, sodic plagioclase and myrmekite and suggest an original source material with a composition between diorite and granite. A small amount of andesite fragments (up to 2%) occurs within these sandstones.

The quartz arenites of the Siegas Formation consist mainly (87 to 95%) of medium grained rounded quartz suggesting a polycyclic history with at least partial derivation from older quartzose sandstones. However, the compositions of the feldspars and lithic fragments of the quartz arenites are identical to those of the arkosic facies, suggesting at least in part a common provenance of felsic plutonic rocks.

The source area of the Siegas Formation is interpreted as a complex local uplift in northwestern New Brunswick consisting of quartzose sandstone, andesite and quartz diorite (Hamilton-Smith, 1971a). This source area would be part of the northern end of the land mass Taconica of Roy (this volume). An analogous association of all three lithotypes is known regionally in the WeeksboroLunksoos Lake anticlinorium (Neuman, 1967).

Siegas Quarry Section

The Siegas Quarry provides an almost completely exposed section through the Siegas Formation in the thickest and most proximal part of the lithic wacke facies. Each bed in the section was numbered, individually described and assigned to one of the lithotypes: limestone conglomerate, sandstone, siltstone, limestone or chert. A first-order Markov process transition matrix was derived in order to define lithological associations. The matrix elements and other features of the five lithotypes are summarized in Table 1.

Table 1: Lithological features of the Siegas Quarry section (after Hamilton-Smith, 1971b).

<u>Lithotype</u>	<u>Transition Matrix</u>					<u>Percent of Thickness</u>	<u>Percent of Beds</u>
	<u>S1</u>	<u>S2</u>	<u>S3</u>	<u>S4</u>	<u>S5</u>		
Conglomerate <u>S1</u>	.27	.68	.05	0	0	13	4
Sandstone <u>S2</u>	.07	.45	.33	.09	.06	72	34
Siltstone <u>S3</u>	.01	.29	.32	.37	.01	12	39
Limestone <u>S4</u>	.01	.15	.68	.15	.01	2	21
Chert <u>S5</u>	0	.75	.17	.08	0	1	2

Inspection of the matrix suggests two empirical lithological associations: an exogenic group consisting of limestone conglomerate, sandstone and chert and an endogenic group consisting of limestone and siltstone. In the section at the quarry the lithotypes of the exogenic group occur in three distinct thickly bedded intervals separated by two thinly bedded intervals consisting of lithotypes of the endogenic group.

The limestone lithotype at the Siegas Quarry is an impure micrite identical to the limestone of the underlying Carys Mills Formation. If a transition matrix is derived for the endogenic group and compared to one from the Carys Mills Formation the results are very similar (Table 2).

Table 2: Transition matrices for the endogenic group, Siegas Quarry section and the Carys Mills Formation (after Hamilton-Smith, 1971b).

<u>Lithotype</u>	<u>Transition Matrix Endogenic Group</u>		<u>Transition Matrix Carys Mills Formation</u>	
	<u>S1</u>	<u>S2</u>	<u>S1</u>	<u>S2</u>
Limestone <u>S1</u>	.23	.77	.15	.85
Siltstone <u>S2</u>	.42	.58	.55	.45

The interpretation of the transition matrices is that the endogenic group of lithotypes records an extension of the sedimentary conditions of the Carys Mills Formation throughout the time of deposition of the Siegas Formation. The exogenic group records three distinct interruptions of this relatively continuous sedimentation by the deposition of sandstone and limestone conglomerate.

Most of the sandstone beds display the classic characteristic features of turbidite deposition. These features can be well observed on the Siegas Quarry and have been fully described previously (Hamilton-Smith, 1970, 1971b). The distribution of sedimentary structures in the section, particularly in the most significant middle exogenic event, indicates a passage from relatively distal to highly proximal and back to distal depositional conditions. An important feature of the Siegas Quarry section is the presence of thick nongraded sandstone beds and subintervals near the base of graded beds associated with the most proximal phase of turbidite deposition. These nongraded beds and intervals contain scattered limestone clasts up to 24 inches in maximum dimension and are interpreted as grain flow deposits.

The limestone conglomerate beds are associated with the turbidite deposition. They generally consist of limestone and slate clasts up to 24 inches in maximum dimension (60% average) and a sandstone matrix compositionally identical to the lithic wacke of the Siegas Quarry. The limestone and slate clasts are lithologically identical to the lithotypes of the endogenic group and the Carys Mills Formation. The angularity, pull-apart structures and plastic deformation of individual limestone clasts suggest a short history of transportation and semi-consolidation at the time of erosion.

Half of the limestone conglomerate beds are strongly graded and have load casts, erosional surfaces and clast imbrication at the bases of the beds. They occur at the most proximal phase of the middle exogenic event and are interpreted as the product of deposition from turbulent flows with unusually high values of effective density and turbulent intensity. Such values may have been produced by lateral confinement of the flow by a narrow channel.

The other limestone conglomerate beds are not graded and have an internally heterogeneous structure. They were probably produced by various forms of mass flow from incoherent slumping to a complex sliding of relatively coherent lenticules over one another with most of the shear stress concentrated near lenticule margins.

The limestone conglomerate beds originated from a mixture of limestone and slate clasts and unconsolidated sand consisting mainly of andesite fragments. The limestone and slate was probably derived from submarine erosion of the Carys Mills Formation a short distance to the north of the Siegas Quarry. Mixing with the sand may well have occurred by collapse of submarine canyon walls into unconsolidated sand forming the channel floor. Such events may also have initiated the high density for turbulent flows that resulted in the deposition of graded limestone conglomerate beds.

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Itinerary

Mileage

- 0 Assembly for trip in Presque Isle, Maine. Starting time 7:45 a.m. Drive north on Route 1 to Van Buren, Maine.
- 33.0 Enter Van Buren, Maine.
- 34.0 Cross St. John River into St. Leonard, New Brunswick. Positive identification of citizenship is required for crossing and return. Drive northeast on Route 17 to intersection with TransCanada Highway.
- 35.3 Turn left onto TransCanada Highway and proceed westbound towards Edmundston.

- 35.9 Stop 1. Park along shoulder of the road well clear of the highway. Outcrop is in the road cut on both sides of the highway.

New Sweden Formation. The exposure is located on the eastern limb of the Ashland Synclinorium and consists of tightly folded light grey, calcereous laminated slate with laminated manganiferous and iron-rich siltstones.

Return to cars. Proceed west on highway.

- 37.8 Stop 2. Park along shoulder of the road well clear of the highway. Outcrop is in the road cut on both sides of the highway.

Jemtland Formation. The outcrop is to the west of the axis of the Ashland Synclinorium just east of the map-area of Figure 1. The outcrop consists of thinly laminated, fissile, calcareous siltstone, shale and sandstone.

Return to cars. Proceed west on highway.

- 40.5 Stop 3. Park along shoulder of the highway well clear of the road. Outcrop is in the road cut on the north side of the highway.

Carys Mills Formation. The exposure consists of steeply dipping interbedded dark grey calcareous slate and limestone with minor sandstone. The limestone beds weather light blue grey or light brown. The sequence occurs near the base of the Carys Mills Formation.

Walk west along the highway to Stop 4.

Stop 4. The outcrop is in the road cut on the south side of the highway. Cross with due regard for the occasionally very fast traffic.

Carys Mills Formation and Siegas Formation. From east to west the outcrop consists of thinly interbedded dark grey calcareous slate and limestone of the Carys Mills Formation succeeded by massive medium grey lithic sandstone beds of the Siegas Formation. The contact is an erosional one, with clear truncation of laminae in the Carys Mills slate and about a foot of exposed relief.

Return to cars and proceed west along the highway.

- 43.2 Stop 5. Park along the shoulder of the highway well clear of the road. The outcrop is in the road cut on the north side of the highway.

Madawaska Lake Formation. The exposure is in the core of the Pennington Mountain anticlinorium. The lithology is dark grey non-calcareous sparsely laminated slate with minor thin light grey sandstone beds.

Return to cars and turn around in order to proceed back down the highway to the east.

- 45.5 Turn left on to access road to the old Highway 2. Immediately turn left again on old Highway 2 and proceed slowly about 150 yards to a gravelled side road. Turn right on the side road and proceed 200 yards to a large open space and park.

Stop 6. The Siegas Quarry. Park in the old processing area at the end of the access road south of the quarry. Walk north to the quarry and proceed to the east end of the exposure at the processing area level.

New Sweden Formation and Siegas Formation. The eastern end of the exposure consists of light grey, calcareous, laminated and locally contorted slate of the New Sweden Formation. The top of the Siegas Formation is defined at the top of the first significant sandstone bed. The contact is conformable. The rest of the exposure consists of almost the entire Siegas Formation in continuous sequence and exposure for about 750 feet of section. The basal beds and the contact with the Cary Mills Formation are obscured but outcrops of the Carys Mills Formation may be observed in the fields immediately to the west of the quarry. The section dips steeply to the northwest and is slightly overturned.

There is an abundance of sedimentary features characteristic of turbidite deposition from both distal to highly proximal phases. Included are graded sandstone and limestone conglomerate beds, massive nongraded beds of sandstone and slump deposits of limestone conglomerate. A wide variety of cross-lamination, convolute lamination and sole features are common.

The Siegas Quarry is the property of Atlas Construction Co. of Fredericton, New Brunswick.

TRIP C-3

BIOSTRATIGRAPHIC TRIP ACROSS NORTHEASTERN MAINE

by

William H. Forbes

This trip will permit the participants to observe and sample fossil localities from Middle Ordovician to Middle Devonian in age across northeastern Maine. The localities are among the best in the eugeosynclinal Appalachians and include outcrops with shelly and/or graptolitic fauna as well as outcrops with abundant Devonian flora.

TRIP C-4

LATE-GLACIAL AND HOLOCENE GEOLOGY
OF THE MIDDLE ST. JOHN RIVER VALLEY

J. Steven Kite and Harold W. Borns, Jr.
James Madison University and University of Maine at Orono

Introduction

The middle St. John River valley lies between the towns of Grand Falls, New Brunswick and Frenchville, Maine (Figure 1). Throughout most of this field area the St. John River serves as the U.S.-Canadian international boundary. Complete topographic coverage of the area is included on three maps: the Canadian (1:50,000) Saint Andre and Edmundston quadrangles and the American (1:62,500) Stockholm quadrangle. The American (1:62,500) Van Buren, Grand Isle, and Frenchville quadrangles include portions of Maine also shown on the two Canadian maps. Most of our field work in this area was undertaken in 1977 and 1978, although this research is currently being extended into the upper St. John River valley.

The glacial history of northern Maine and New Brunswick greatly influenced late-glacial and Holocene events. Ongoing studies on both sides of the international border are putting together the Late Wisconsin glacial history of the region (Genes and Newman, Trip B-8; Gauthier, Trip B-9). The chronology determined by these studies undoubtedly will add to our understanding of postglacial events; however, in light of current knowledge we believe two glacial events, both recognized at least three decades ago, controlled drainage in the St. John watershed. The first event was the formation of an ice dam in the Edmundston area which created Lake Ennemond at altitudes of up to 680 ft in the western end of the field area (Kiewiet de Jonge, 1951). The second important glacial event was the construction of a large moraine across the valley at Grand Falls; after deglaciation, this moraine dammed Lake Madawaska, which extended throughout the middle St. John River valley at altitudes of up to 513 ft (Chalmers, 1886; Kiewiet de Jonge, 1951; Lee, 1963).

Lake Madawaska existed until about 10,000 B.P. when water level rapidly declined. The emergent lake bottom was covered first by extensive early Holocene swamps, and later by three different facies of early-to-middle Holocene alluvium. Two of these alluvial facies have been misinterpreted in other studies as either glaciolacustrine or glaciofluvial deposits even though identical sediments are accumulating in present-day St. John River environments. The St. John River has incised its bed in late Holocene time, leaving the old fluvial and lacustrine deposits exposed in terrace scarps.

Acknowledgements

Robert Stuckenrath of the Smithsonian Institution provided 11 new radiocarbon dates and aided in the interpretation of these dates. A

Grant-in-Aid of Research from Sigma Xi, and Fellowships from the University of Maine at Orono supported field work. Keith A. Laskowski and Susan C. Kite provided able field assistance. This work benefited greatly from discussions with Andrew Genes, Glenn Prescott, George Denton, Barry Timson, Thomas Lowell, Bradford Hall, and Stephen Norton.

Stratigraphy and Geomorphology

Ice-contact stratified drift is the lowest Quaternary strata exposed in the middle St. John River valley. This drift occurs in kames, kame terraces, and the moraine at Grand Falls. The stratified drift is composed of clasts ranging from fine silt to cobble-boulder gravel. Canadian Shield gneissic erratics are a minor constituent. Collapse features such as normal faults and steeply dipping beds are common. Flow till occurs at several localities. In most exposures, ice-contact stratified drift is capped by a layer of till; in one gravel pit the stratified drift underlying the till layer exhibits overturned glaciotectionic folding (Stop 10). The stratified drift presumably accumulated during the next-to-last glaciation of the valley, the overlying till dates to the last glaciation. Several isolated ice-contact deposits in the field area are not capped by till and are probably younger than the till-mantled stratified deposits.

There are two distinctive tills which mantle the bedrock hills, kames, and kame terraces in the field area. A well-compacted grey (N4: wet color) diamicton is probably a lodgement (basal) till. This grey till is extremely poorly sorted; clasts are predominantly limestone and fine-grained detrital rocks with minor constituents of conglomerate, gneiss, and volcanic rocks. Up to 90 percent of cobble and boulder clasts are striated and there is a distinct orientation of elongate clasts. An oxidized greyish orange (10 YR 7/4) to yellowish brown (10 YR 5/2) till, less compact than the grey till, is widespread throughout the field area. This oxidized till was deposited as ablation or flow till. Again, gneissic erratics are minor constituents. Bedding is generally absent, but rare layers of sand or silt occur. The oxidized till has less distinctive fabric and fewer striated clasts than the well-compacted grey till. Despite distinct differences between the two tills, we do not believe there is sufficient evidence within the middle St. John River valley to assign these tills to different time periods or different ice masses. Both tills occupy the same stratigraphic position overlying the majority of ice-contact stratified drift in the field area; hence, we believe the tills represent different facies of the last glaciation. Older tills undoubtedly underlie the extensive ice-contact stratified drift, but we have not recognized outcrops of older till in this field area.

Two outwash deposits, composed of rounded cobbles and boulders with a sand-and-silt matrix, exist in the field area. A valley fill composed of up to 200 ft of rounded gravel lies at the eastern end of the field area. The head of this outwash train adjoins the Grand Falls moraine. South of Grand Falls, the outwash occurs in a steep-scarped terrace which is over 20 miles long. A second valley fill, 25 ft thick, is exposed in

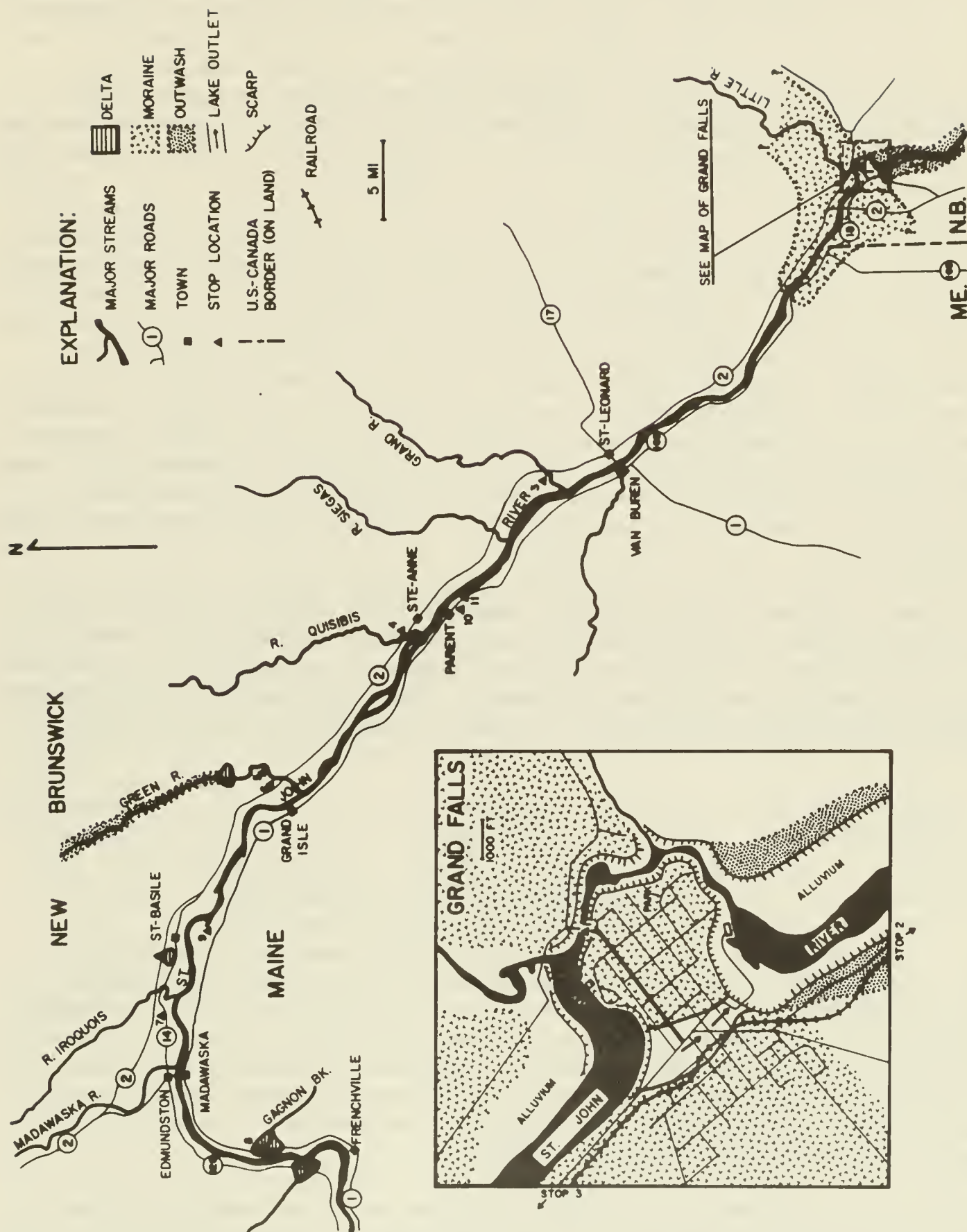


Figure 1. Middle St. John River valley. The St. John River is the international boundary throughout most of the field area.

terraces along the Green River valley. The outwash head is a small moraine north of the field area (Lee, 1955). Gneissic erratics occur at both localities.

Rhythmically bedded, grey (N5) silt and clay make up a large portion of the postglacial subsurface deposits in the field area. However these lacustrine rhythmites occur on the valley floor below an altitude of 505 ft and are not widely exposed because of overlying alluvium. Each rhythmic bed is 0.1 to 1.0 in thick and displays graded bedding. The lack of sand-silt rhythmites or thick basal rhythmites indicate these beds accumulated in deep water, removed from active ice margins or deltas. Rhythmic bedding thins and becomes less distinctive upward, uppermost beds are massive.

We have mapped two types of lake-margin deposits in the field area: beach ridges and deltas. Beach ridges, generally less than 5 ft high and up to 600 ft long, are composed of well-sorted gravel and sand. All of the known beaches are associated with Lake Madawaska strands lying between altitudes of 463 ft and 513 ft. Unlike beaches, deltaic deposits are associated with both the highest strand of Lake Madawaska (about 510 ft) and with several Lake Ennemond water levels (580 ft to 680 ft). Deltaic deposits closely resemble outwash sediments, except that deltas exhibit steeply dipping foreset bedding, which is not found in outwash. One Lake Madawaska delta lies at the mouth of the Green River valley outwash, suggesting that glacial ice near the field area co-existed with the earliest lake strand.

Buried peat beds are exposed at two outcrops (Stop 5 and Stop 11). Plant fragments are humified in the lower portion of both peat beds, whereas the upper portion contains wood fragments up to 4 in in diameter. The peat probably accumulated in a swamp that formed over Lake Madawaska sediments after water level fell. At both localities, the peat beds underlie alluvium of the highest alluvial terrace.

Channel-lag and channel-bar sediments are the coarsest of three distinct alluvial facies that can be identified in both modern-day sediments and Holocene strata. These channel sediments, ranging from coarse sand to cobble-gravel, commonly are 10 to 12 ft thick exhibiting imbrication, crude horizontal bedding, and high-angle cross-bedding. In one exposure, channel gravel is indurated into a conglomerate with limonite cement. Channel-lag and channel-bar alluvium can be distinguished from outwash using three criteria: 1) outwash deposits must have an outwash head; 2) coarse-grained deposits may be attributed to adjacent stream deposition, if the present day stream is capable of transporting comparably sized sediments; and 3) it is unlikely that the glaciers necessary to generate outwash remained in the field area long after the climatic amelioration that occurred about 10,000 B.P.

The coarse-grained channel deposits are overlain locally by grey (N5) channel-fill silt and clay, widely exposed along the banks of the St. John River and its tributaries. The fine-grained channel-fill lenses range from 3 to 15 ft in thickness and some exceed one-half mile in length. They commonly parallel the present-day stream course giving rise to outcrops that are thousands of feet long. Bedding appears massive,

except in horizons rich in flat-lying plant fragments. Limonite mottling and amorphous blue masses of vivianite commonly surround plant fragments. We have collected numerous flattened leathery mollusk shells, which have been leached of calcite. Present-day channel-fill basins (such as the basin seen at Stop 9) are subaerially exposed only during low stages of the St. John River.

Most of the alluvium cropping out in the area is floodplain sand and silt, which may be up to 23 ft thick. Reverse grading of individual beds is common, although sequences of beds deposited in a floodplain environment become progressively finer-grained toward the top of the sequence. Low-angle cross-bedding occurs in fine sand and coarse silt; angle-of-repose cross-bedding exists in coarse sand beds. Bedding cannot be discerned in weathered silt near the top of alluvial terraces upon which floodplain deposition no longer occurs. Floodplain alluvium is the uppermost (hence, youngest) strata in every exposure in the field area.

The three alluvial facies discussed above make up the highest alluvial terrace in the St. John River valley. This highest terrace is best exposed in the vicinity of Siegas, New Brunswick; hence we call the landform the Siegas terrace. The Siegas terrace can be traced throughout most of the reach between Grand Falls and Edmundston, at altitudes of 33 to 40 ft above low stage of the St. John River. The gradient of the terrace surface is approximately the same as the gradient of the St. John River prior to construction of the hydroelectric dam at Grand Falls. All three alluvial sediments also occur in lower terraces, which lie 25 to 33 ft above low river level. These lower terraces are not paired and probably were formed during progressive downcutting of the St. John River channel. The present-day alluvium also includes the three alluvial facies. Channel-lag, channel-bar, and channel-fill sediments are restricted to low-lying areas near streams. Active floodplain accumulation, evidenced by lack of soil profile development, occurs up to 25 ft above low stage of the St. John.

Late-Glacial and Holocene Geologic History

The middle St. John River valley was deglaciated in stages. After the deglaciation of the St. Lawrence lowlands at about 13,000 B.P., a residual ice cap remained in northern Maine and New Brunswick (LaSalle, et al., 1977). Cut off from ice domes in the Canadian Shield, this ice cap thinned and broke into smaller glaciers. One of these glaciers stood between Edmundston and Grand Falls while the western and eastern ends of the field area were ice-free. This glacier probably dammed Lake Ennemond and deposited the moraine and extensive outwash at Grand Falls. During the highest strand (680 ft) of Lake Ennemond, the lake may have drained through one of several potential outlets south of the field area (Figure 2A). Subsequent lower strands (660 to 580 ft) may record the opening of an outlet through the middle St. John River valley as the glacier wasted in place. It is possible that an arm of Lake Ennemond extended east of Edmundston as the glacier contracted. After the glacier melted completely,

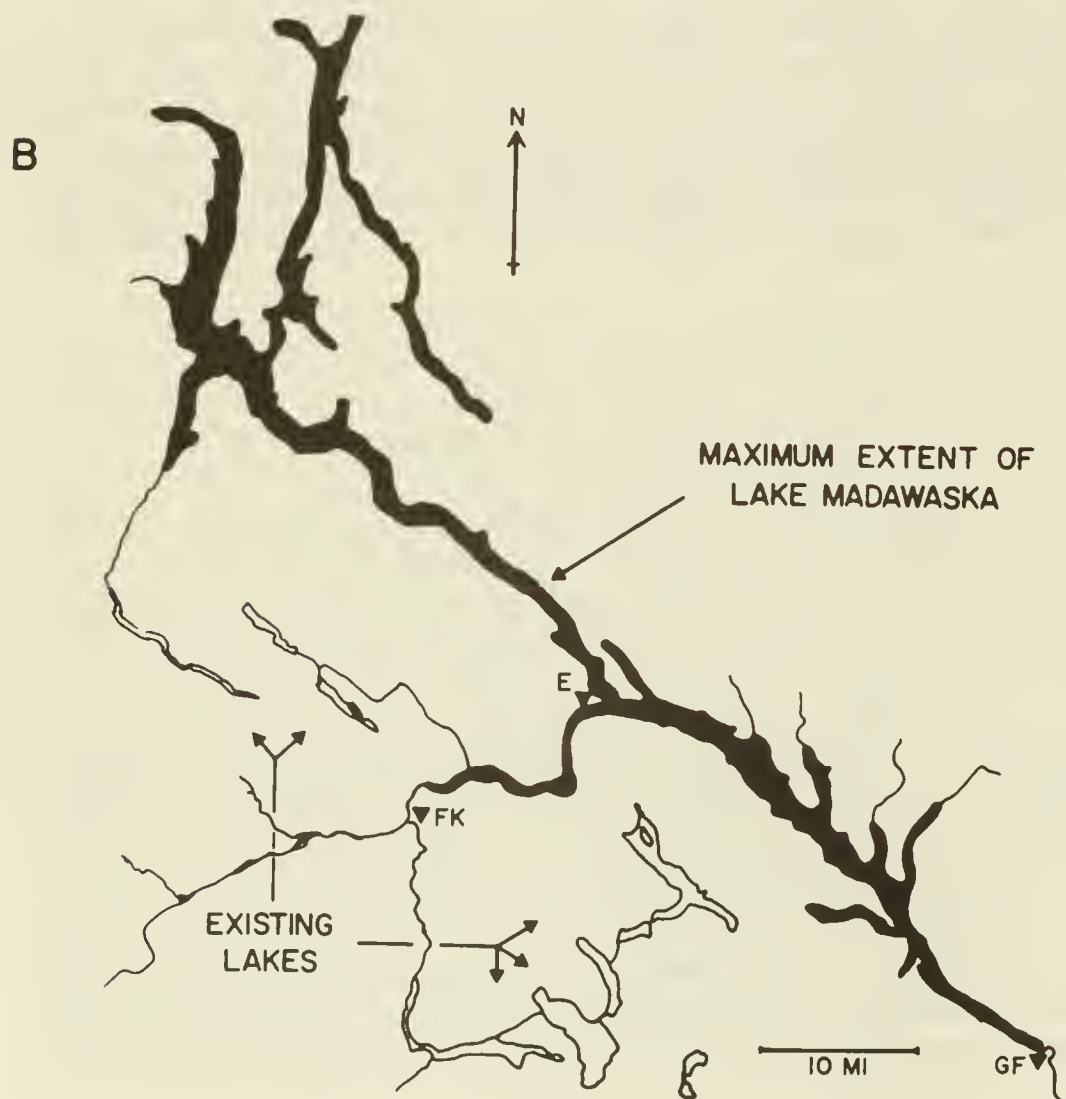
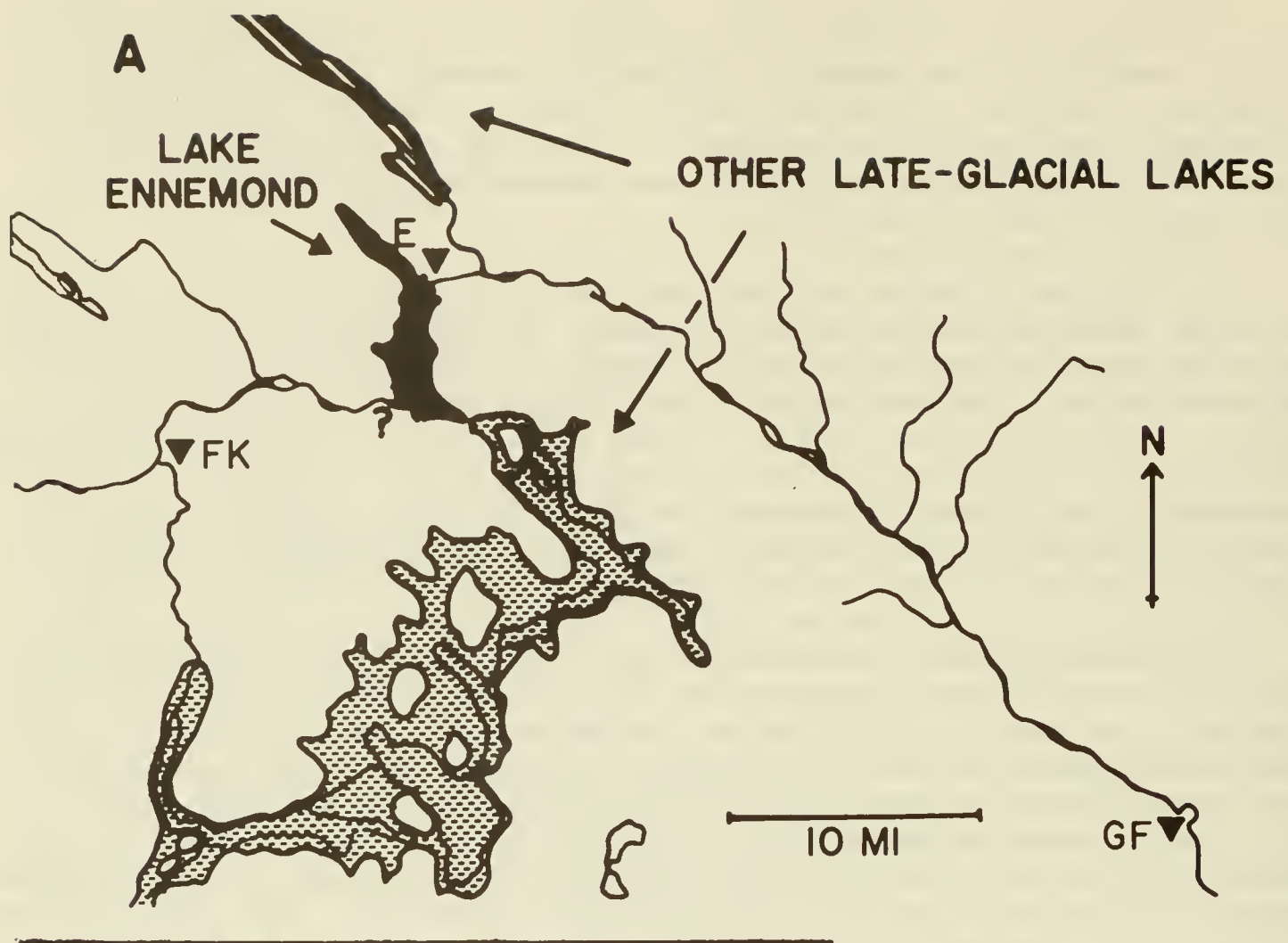


Figure 2. Ancient lakes of the middle St. John River valley: A) maximum extent of Lake Ennemond and B) maximum extent of Lake Madawaska. Towns are denoted with triangles: GF = Grand Falls, E = Edmundston, FK = Fort Kent.

the only obstruction of drainage in the valley was the Grand Falls moraine and associated outwash, the drift which dammed Lake Madawaska (Figure 2B). In its early history the lake apparently drained through two outlet channels at Grand Falls. One outlet paralleled the present-day course of the St. John, whereas the other outlet was about one-half mile south of the Grand Falls gorge (Figure 1).

The date of deglaciation and onset of lacustrine sedimentation can be inferred from an exposure on Green River (Stop 5); here, an estimated 1200 rhythmites underlie the base of a peat bed which has yielded 5 credible radiocarbon dates of about 10,000 B.P., with an average of 9979 ± 108 B.P. If the rhythmites are assumed to be varves, and if we assume there is no hiatus in the stratigraphic sequence, then the oldest lake sediments accumulated at about 11,200 B.P. Deglaciation probably occurred immediately before deposition of the lake sediments. The highest recognized strand deposits in the vicinity of the Green River exposure occur at elevations of about 510 ft; the altitude of these strands suggests that the 11,200 B.P.-aged lake sediments accumulated in Lake Madawaska.

Beaches below the highest strand record water levels during the regression of Lake Madawaska (Figure 3). The only well-exposed beach in the field area (Stop 10) overlies a shallow-water silt deposited while other beaches formed at higher altitudes. As the water level fell, the emergent lake bottom was covered by swamps. The 10,000 B.P.-aged peat bed at the Green River exposure (altitude: 474 ft) formed after Lake Madawaska dropped about 36 ft below its highest nearby strand. This 36 ft regression occurred in about 1200 years, indicating an average water level drop of 3 ft/100 years. A 9,900 B.P.-aged peat bed near Parent, Maine (altitude: 446 ft; Stop 11) suggests that the lake level fell 28 ft within the century after 10,000 B.P. Indeed, the radiometric control over the Parent and Green River exposures does not preclude the possibility that the 28 ft drop in lake level occurred catastrophically. Our field data suggest the decline after 10,000 B.P. was so rapid that no mappable beach deposits formed below the 10,000 B.P. strand. The southern outlet to Lake Madawaska was probably abandoned during the rapid drop in lake level. Organic-rich channel-fill sediments taken from the abandoned outlet yielded a radiocarbon date of 9830 ± 160 B.P. (GSC-56).

We have considered a number of possible causes for the rapid demise of Lake Madawaska. Three of the more probable causes include isostatic uplift in the upper St. John watershed, failure of a drift dam or ice dam, and changes in the hydrology of the watershed induced by climatic amelioration. Research in the upper St. John valley hints that other extinct lakes in the watershed had similar histories to Lake Madawaska.

Wood incorporated into channel gravel indicates that a river flowed over the emerged lake bottom by 9820 ± 130 B.P. (GSC-18), or possibly somewhat later if the wood was reworked prior to deposition. Lake Madawaska ceased to exist before this event, although small discontinuous lakes may have persisted for a short time in lower areas of the valley. The early Holocene St. John River deposited gravel at altitudes as low

LAKE MADAWASKA STRANDLINES

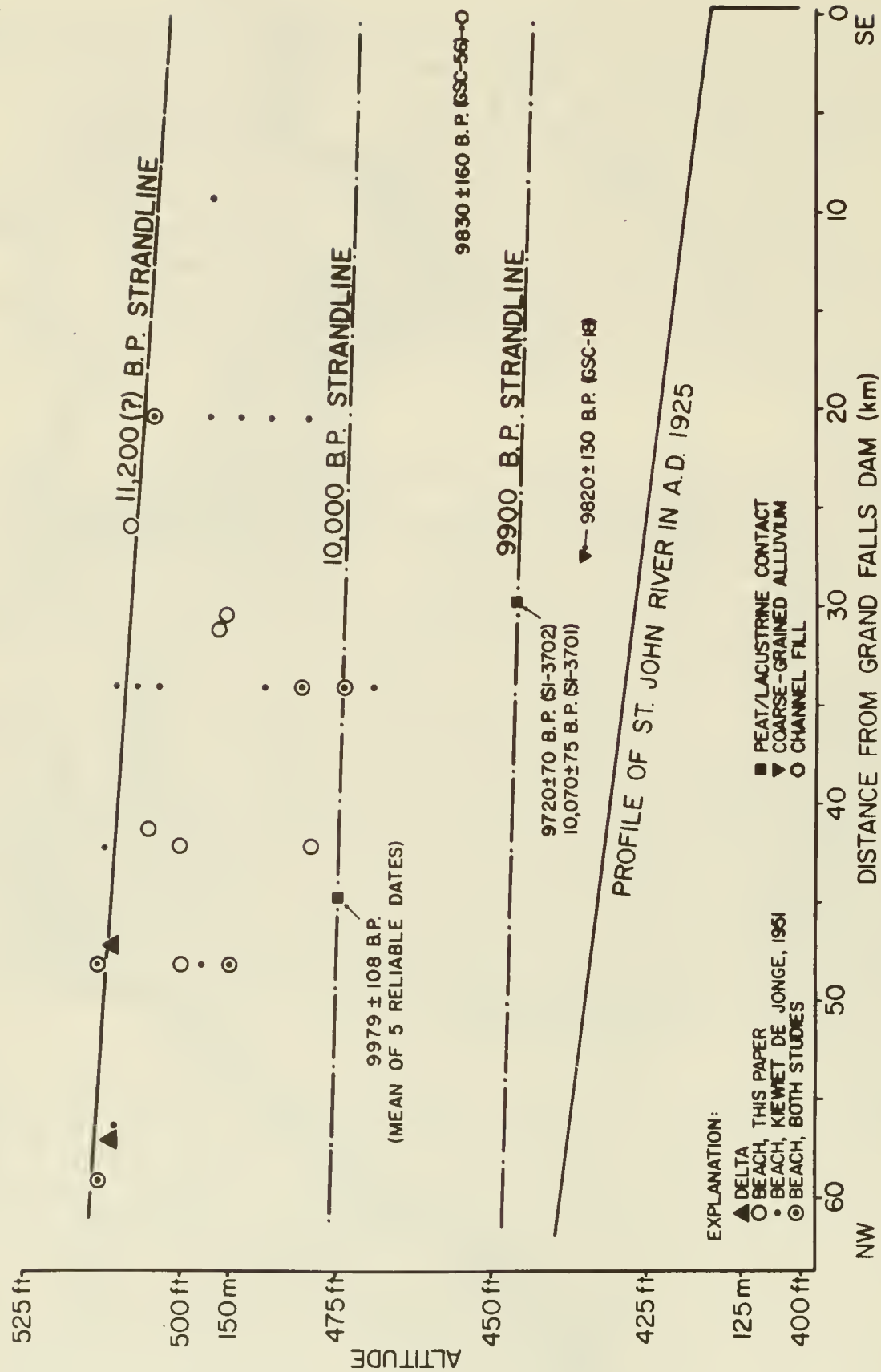


Figure 3. Lake Madawaska strandlines. The 10,000 B.P. and 9900 B.P. strandlines are inferred from age of peat/lacustrine contacts; tilt of both strandlines is based on arbitrary assumption that isostatic warping was one-half of the warping in the 11,200 (?) B.P. strandline.

as 427 ft. The newly formed reach of the river had an extremely low gradient over the former lake bottom. Immediately after the draining of the lake the St. John River and its tributaries began an aggradational episode, which was widespread by 8200 B.P. (Figure 4). Aggradation continued until after 4250 ± 50 B.P. (SI-3703). By the end of the aggradational episode a broad floodplain existed with a surface that sloped from 480 ft near Grand Isle to 458 ft near Grand Falls. Small, coarse-grained alluvial fans, a few feet above the floodplain surface, formed near the mouths of many tributaries.

Downcutting of the St. John River began once the river attained a graded profile. The floodplain and alluvial fans were incised to form the Siegas terrace. Lower terrace-like surfaces fall into two age groups. Weathered surfaces lying 5 to 15 ft below the Siegas terrace are alluvial terraces abandoned during the late Holocene. Unweathered surfaces 10 to 30 ft below the Siegas terrace are part of the present-day floodplain.

Geoarcheology

We did not find any artifacts in our field work, but the stratigraphy of the middle St. John River valley suggests potential for significant archeological sites. Lake Madawaska persisted long after the arrival of man in Maine and the Maritimes, as evidenced by the Paleo-Indian site at Debert, Nova Scotia, which is dated $10,589 \pm 47$ B.P. (Stuckenrath, oral communication, 1979). Some of the Lake Madawaska beaches apparently are contemporaneous with the Debert site, and significant Paleo-Indian sites may be located on, or just above, these relict lake shores. The alluvial stratigraphy of the Siegas terrace spans most of the period between 10,000 B.P. and 5000 B.P., during which time there is little known about man in Maine and the Maritimes. The Siegas terrace probably contains sites that could fill this void in the archeological record. Sites of this age within the Siegas terrace may have a well-developed stratigraphy in which each artifact assemblage is temporally separated from other assemblages. Late Holocene artifacts probably occur in surface sites on the Siegas terrace, surface and buried sites on younger alluvial terraces, and buried sites on the present-day floodplain.

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EARLY TO MIDDLE HOLOCENE AGGRADATION

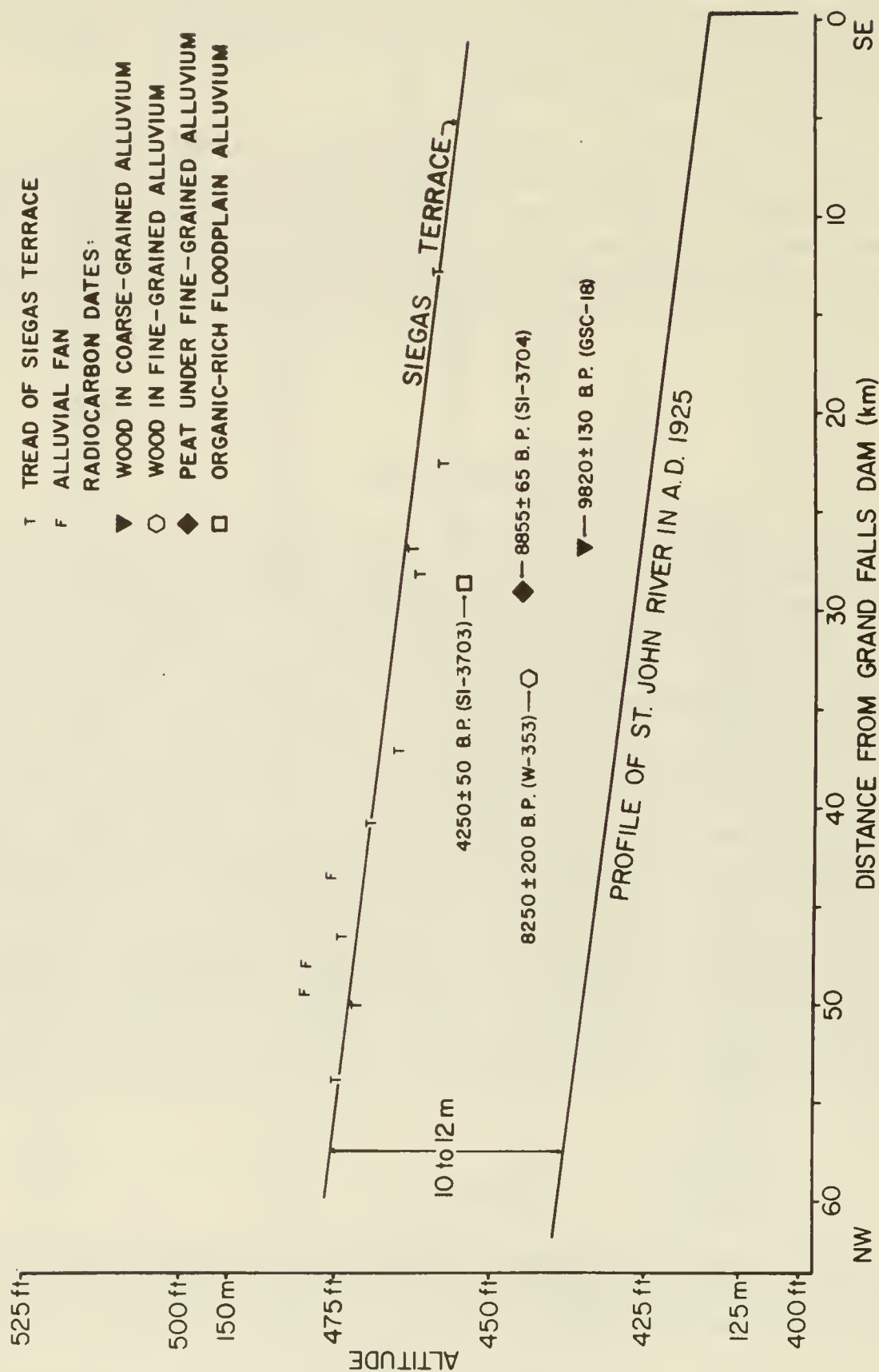


Figure 4. Early to Middle Holocene aggradation. Three other dates were obtained from the organic-rich floodplain alluvium: 5795 ± 70 B.P. (SI-3901), 6160 ± 85 B.P. (SI-3900), and 8770 ± 100 B.P. (SI-3901A).

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Lee, H. A., 1955, Surficial geology of Edmundston, Madawaska and Temiscouata Counties, New Brunswick and Quebec: Geological Survey of Canada Paper 55-15, 14 p.

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Itinerary

Mileage

0.0 Assemble at entrance to Centennial (Municipal) Park, Grand Falls, New Brunswick, Saint-Andre 1:50,000 quadrangle. Starting time 8:30 A.M. Eastern (9:30 A.M. Atlantic). Walk west 300 yd to bridge over St. John River.

0.0 Stop 1. Walk to center of bridge, facing waterfall.

Grand Falls at the confluence of St. John and Little Rivers. The St. John drops over 120 ft within one mile of the hydroelectric dam; most of the drop occurs at this waterfall. The view is most spectacular during April or May, when spring melt-water increases discharge to as much as 1.65 million U.S. gallons per second. Carefully walk to the other side of the bridge to view gorge. The erosive power of the river during peak discharge is evidenced by the large potholes and the treeless zone that extends far up the sides of the gorge. Before 10,000 B.P., the path of the gorge was the northernmost of two Lake Madawaska outlets. The southern outlet, 900 yd from this bridge, was underlain by over 200 ft of drift in the Grand Falls moraine (Figure 1). The drift-floored southern outlet was abandoned about 10,000 B.P. The northern outlet probably was drift floored at 10,000 B.P., also; if it had been bedrock floored, the St. John River would have become entrenched in the more easily eroded southern outlet. Hence, the deep gorge beneath the bridge was apparently carved after 10,000 B.P. because of superposition of Lake Madawaska's northern outlet.

Return to assembly point, proceed in vehicles northwest on Victoria St.

0.2 Turn left on Broadway.

0.6 Turn left on Condon St. The railroad tracks visible south of the intersection overlie the abandoned southern outlet of Lake Madawaska.

- 0.8 Cross railroad tracks.
- 0.95 Turn left on River Road at 3-pronged intersection; cross railroad tracks again.
- 2.35 Turn right at entrance to large gravel pit.
- 2.35 Stop 2. Park well out of truck traffic.

Grand Falls outwash. This coarse gravel was deposited by meltwater from the glacier that built the Grand Falls moraine. Cobble and boulder lithologies include Canadian Shield gneisses which originated west of the St. Lawrence, it is possible that these erratics were not transported across the St. Lawrence by the Grand Falls glacier but were reworked from older drift. Large trough cross-bedding was exposed in the southern face of the pit in 1977 and 1978. A veneer of sand and silt caps the coarse gravel. These finer sediments probably mark a reduction in meltwater after the glacier retreated from Grand Falls.

Return to vehicles, turn left on River Road.

- 3.75 Turn right on Condon St., immediately after crossing railroad tracks.
- 5.2 Turn left onto Route 2 (Trans-Canada Highway) west.
- 9.4 Bridge over Canadian Pacific Railroad tracks; note the view to southwest across St. John River. Landforms include modern-day floodplain, the Siegas alluvial terrace (site of several farms and paved road) and a till-mantled bedrock ridge. These landforms are shown on both Saint-Andre 1:50,000 and Van Buren 1:62,500 quadrangles.
- 11.6 Note the well-developed segment of the Siegas terrace on this side of the St. John River.
- 16.5 Intersection of Route 2 and Route 17, continue west on Route 2.
- 18.4 Cross Grande Rivière (Grand River).
- 18.5 Stop 3. Park in front of gate on left side of road. Follow trail to river bank 300 yd south of gate.

Three alluvial facies of a late Holocene terrace. At least 6.0 ft of gravel crop out at the base of this exposure. The gravel unit displays sedimentary structures (crude bedding and imbrication) and grain-size distribution similar to present-day channel-lag and channel-bar deposits; the gravel in the exposure probably accumulated in a channel facies. 4.6 ft of massive silt and clay overlie the gravel. Sedimentary characteristics include abundant carbonized plant fragments concentrated near bottom of unit and limonite mottling. The depositional environment was probably similar to the modern-day channel-fill facies.

The channel-fill clay and silt are covered by 3.6 ft of sand and silt which were deposited in the floodplain facies.

Return to cars. Proceed west on Route 2.

21.0 Cross Rivière Siegas (Siegas River).

21.2 Note the unpaved road to left which leads to the abandoned site of Siegas Station. The Siegas terrace, the highest alluvial terrace in the field area, is best developed in this vicinity. An excellent exposure of the three alluvial facies of the Siegas terrace occurs on the Canadian bank of the St. John River, 400 yd southwest of Siegas Station. This exposure can be seen from Stop 11 and from U.S. Route 1 in Maine. The deposits are similar to those exposed at Stop 4. We will not have time to visit this exposure; continue west on Route 2.

24.7 Enter Edmundston 1:50,000 quadrangle in the town of Ste-Anne-de-Madawaska.

25.2 Stop 4. Turn right at last intersection before bridge over Rivière Quisibis (Quisibis River). Follow right turn with a quick left. Park car on dead end road paralleling Route 2. Walk across bridge and proceed to exposure on bank of Rivière Quisibis 125 yd southwest of bridge.

Alluvial facies of the late to middle Holocene Siegas terrace. Again channel gravel, channel-fill silt and clay, and floodplain sand and silt are exposed. The contact between the channel-fill and floodplain facies is marked by alternating beds of sand and fine silt. An organic-rich zone at the base of the channel-fill alluvium yielded wood dated 8250 ± 200 B.P. (W-353, Lee, 1955). During low-water levels, silts, possibly deposited in Lake Madawaska, are exposed below the channel-gravel unit. Another exposure of the same units occurs on the bank of the St. John River 400 yd southeast of this locality.

Return to vehicles, continue west on Route 2.

25.3 Cross Rivière Quisibis.

31.0 Cross Green River (Rivière Verte).

31.2 Turn right on gravel road located a few yards east of exit to town of Rivière Verte (Green River Station). Continue north-east between brick house and campsites; proceed as far as road conditions permit.

32.0 Stop 5. Park before reaching shallow gravel pits; walk to first gravel pit.

Holocene point-bar and floodplain deposits. This operation is extracting gravel that is nearly identical to point-bar

sediments accreting along Green River (Stop 6). Much of the gravel appears to be derived from an outwash plain and a delta (altitude: approx. 510-515 ft) formed at the mouth of the outwash plain north of this locality.

Walk southeast under power line. Cross Green River in vicinity of mid-channel bar. (During high water, fording the stream can be avoided by approaching this outcrop from the east along the power line right of way. However, this eastern approach requires a one-half mile foot traverse through poorly-drained terrain.)

Green River exposure (Figure 5). This is the best exposure of Lake Madawaska sediments in the middle St. John River valley. The compact grey diamicton at the bottom of the exposure displays a distinct orientation of elongate clasts: long axes trend N 60° W. Canadian Shield gneisses are a minor constituent of this diamicton which is probably a basal till. The compact grey till is overlain by a highly deformed unit that includes gravel, sand, silt, and clay. Portions of this unit contain rhythmic beds indicating glaciolacustrine deposition. Glaciolacustrine deposition of the deformed unit is also supported by the fact that this unit underlies a thick sequence of rhythmically-bedded clays and silts which were deposited in Lake Madawaska. There are approximately 1200 rhythmic beds in this exposure. The upper one-half foot of lake sediments does not display rhythmic bedding, possibly because they were deposited in shallow water. Assuming the rhythmic beds are varves (i.e. deposited annually) Lake Madawaska sedimentation lasted for about 1200 years at this locality. The oldest beds of this lacustrine unit are probably the same age as the highest lacustrine beaches and deltas in this vicinity (altitude: approx. 510 ft). The base of the peat overlying the lake sediments has yielded five credible ^{14}C dates with a mean age of 9979 ± 108 B.P. and a median age of $10,090 \pm 130$ B.P. If there was no hiatus between lacustrine and peat deposition the oldest lake sediments date to about 11,200 B.P. Hence, this locality was probably deglaciated shortly before 11,200 B.P. The top of the peat bed dates to 9285 ± 70 B.P. (SI-3706); after this date the peat was buried under the fining upward alluvial sequence exposed at the top of the outcrop.

Return to vehicles, turn around and drive toward Route 2.

32.7 Stop 6. Park in front of campsites, before reaching Route 2. Walk 250 yards southeast to west bank of Green River.

Present-day point-bar and floodplain deposits. The west bank of Green River consists of a point bar. Note the coarse grain size and clast imbrication. Total thickness of the gravel, including gravel below river level, is about 20 ft.

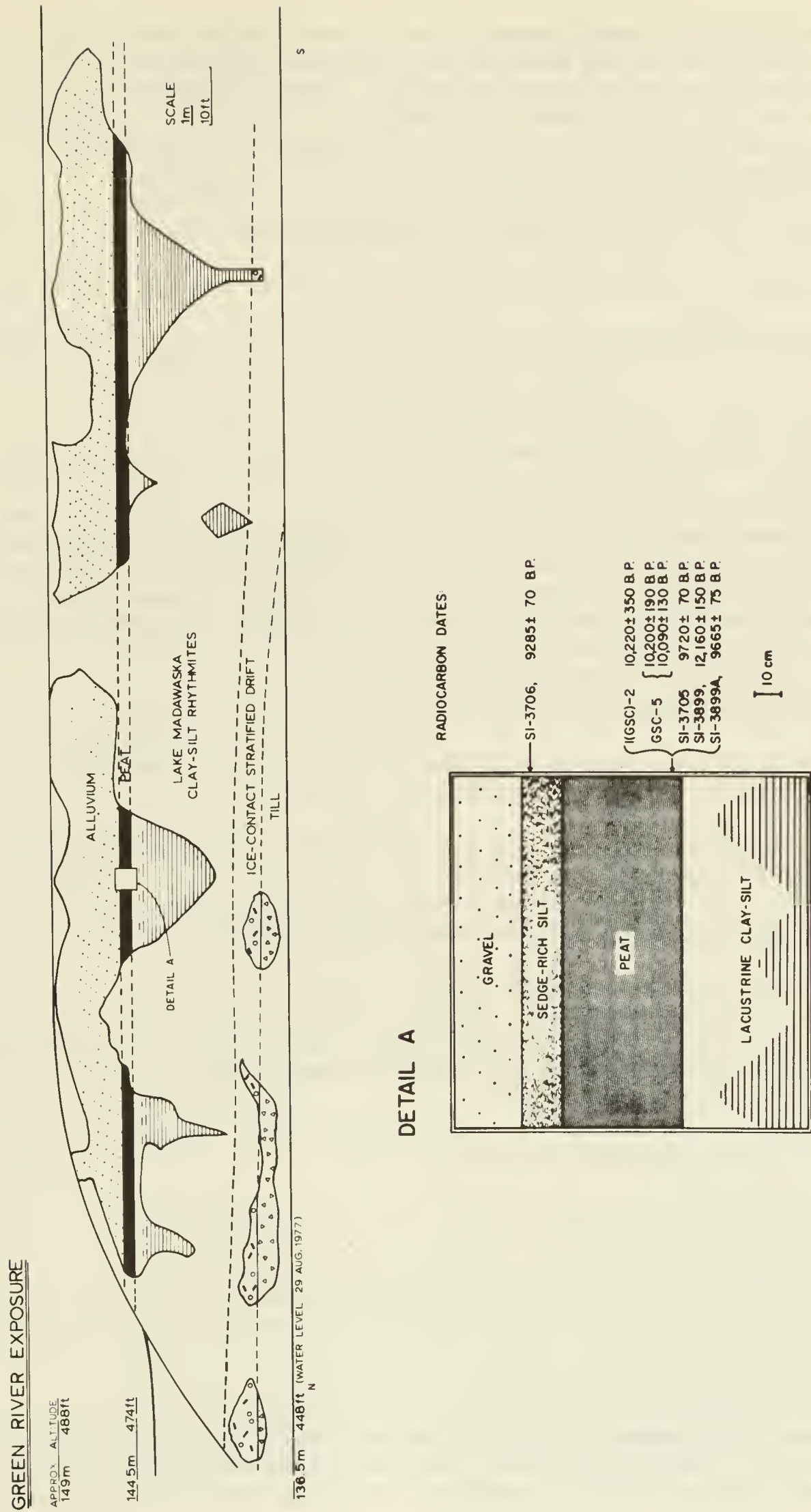


Figure 5. Stop 5: Green River exposure. Five of six radiocarbon dates obtained from samples taken at the bottom of the peat bed cluster around 10,000 B.P. The sample that yielded the anomalous date of 12,160 ± 150 B.P. (SI-3899) probably included reworked organic material derived from older deposits.

Cross bridge to east bank of Green River, north of Route 2. Examine the sand and silt of the floodplain facies. Nearly every bed displays reverse-grading: within each bed the grain size is silt at the base and gradually increases to sand at the top. The contact between individual beds is abrupt. The mechanism producing reverse-graded bedding is unclear.

Return to vehicles, continue toward Route 2.

32.8 Turn right (west) on Route 2.

37.5 Turn left on Route 14.

41.0 Cross Rivière Iroquois (Iroquois River).

41.5 Stop 7. Park on right side of Route 14, north of radio tower.

Lake Madawaska beach. A low ridge in the lawn north of Route 2 is one of the highest beach deposits known between Edmundston and Grand Falls. The beach has an altitude of 512 ft and was probably formed at the same time as the delta along Rivière Verte, early in the history of Lake Madawaska. A cross section of a beach ridge is exposed at Stop 10.

Continue west on Route 14.

42.6 Route 14 passes over an abandoned channel of the Madawaska River.

43.3 Cross Madawaska River; look for signs giving directions to Madawaska, Maine or to Customs.

43.5 Following directional signs bear right, then bear left.

43.6 Turn left on Rue de St. Francois.

43.7 Turn right onto International Bridge; enter Frenchville 1:62,500 quadrangle (this leg is also shown on Edmundston 1:50,000 quadrangle).

44.0 Stop at U.S. Customs, Madawaska office.

44.1 Turn right on U.S. 1.

48.3 Cross Gagnon Brook.

48.6 Stop 8. Turn left into Madawaska Sand and Gravel pit, park well out of truck traffic.

Gagnon Brook delta. This compound delta was built into ice-dammed Lake Ennemond. Several water levels ranging from 680 ft to 580 ft are recorded here, the best developed delta topsets occur at about 620 ft. Relict ice-wedge casts may be exposed in

these topset beds. The Ruisseau Felix-Martin (Felix Martin Brook) compound delta can be seen in New Brunswick, one mile southwest of this locality.

Return to Route 1, turn right.

58.6 Enter Grand Isle 1:62,500 quadrangle (also shown on Edmundston 1:50,000 quadrangle).

59.3 Turn left on unpaved road.

59.6 Stop 9. Park on top of terrace; do not attempt to drive down terrace scarp. Walk to first culvert.

Cyr's Ponds: abandoned channel sediments. The silt and clay accumulating in these water bodies are analogous to the massive silt and clay of the channel-fill deposits. Comparison of grain-size distribution, sedimentary structures (especially massive bedding and limonite mottling), and faunal assemblage shows the sediments accumulating in these basins are nearly identical to those exposed at Stops 3, 4 and 11. The abandoned channels are not true ponds but are backwater basins in which water level is controlled by the height of the St. John River. The channel-fill sediments are exposed only during low water levels. Floodplain sands and silts accumulate on higher surfaces nearby.

Return to vehicles, turn around and drive to Route 1.

59.9 Turn left on Route 1.

70.9 Enter Stockholm 1:62,500 quadrangle (this area is not shown on any 1:50,000 Canadian quadrangle).

72.8 Note unpaved road on right, immediately before passing over knoll east of Parent, Maine. Continue southeast on Route 1.

72.9 Turn right on next unpaved road, just east of knoll crest.

73.0 Stop 10. Park near entrance to gravel pit. Walk to southwest end of pit.

Till-mantled kame underlying beach built during regression of Lake Madawaska. The southwest face exposes ice-contact stratified drift which underlies an oxidized buff till. Glaciotectionic isoclinal folding was exposed in 1977 near top of ice-contact stratified drift. One to two feet of shallow water lacustrine silt covers the till. At the southern corner of the pit the silt underlies a beach ridge composed of subrounded gravel and sand. The beach crest has been stripped away. The existence of higher beaches nearby, coupled with the presence of shallow-water silt under the beach gravel demonstrate that this beach formed during regression of Lake Madawaska.

Return to Route 1.

73.1 Turn right on Route 1.

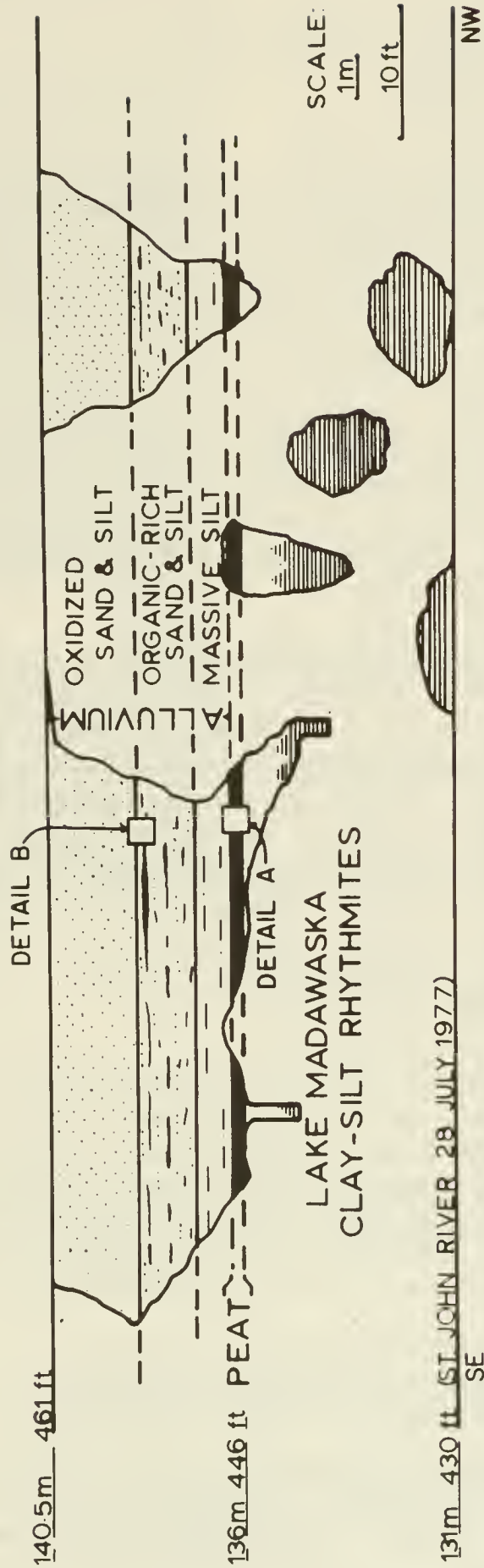
73.8 Stop 11. Park on right side of road across from white transformer. Walk 100 yd northeast to St. John River bank.

Parent exposure (Figure 6). At least 700 rhythmic beds of Lake Madawaska silt and clay are exposed near river level; the basal contact of this lacustrine unit is not exposed. Radiocarbon dates obtained from samples taken along contact between lacustrine sediments and overlying peat bed indicate that the level of Lake Madawaska dropped below this elevation (446 ft) about 9900 B.P. Peat deposition ended about 8855 ± 65 B.P. (SI-3704). The massive grey silt and clay overlying the peat bed is similar to sediments previously identified as channel-fill deposits. However, these massive sediments are not underlain by channel gravels. These sediments probably accumulated in a flood basin that formed when the St. John River began building levees during the early Holocene aggradation. The uppermost horizons in the exposure are floodplain deposits. The youngest radiocarbon date obtained from organic lenses in the floodplain sediments is 4250 ± 50 B.P. (SI-3703). Floodplain deposition on the upper surface of the Siegas terrace ended after this date.

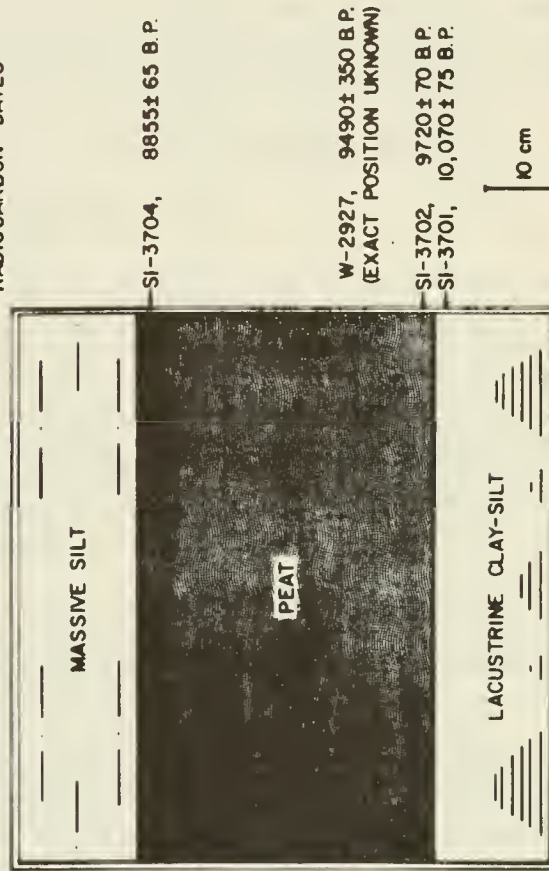
Return to vehicles. End of field trip. Those returning to Presque Isle should continue southeast on Route 1, taking care not to miss the right turn in Van Buren.

PARENT EXPOSURE

(GENERALIZED FROM 0.5 km OF EXPOSURE)



DETAIL A



DETAIL B

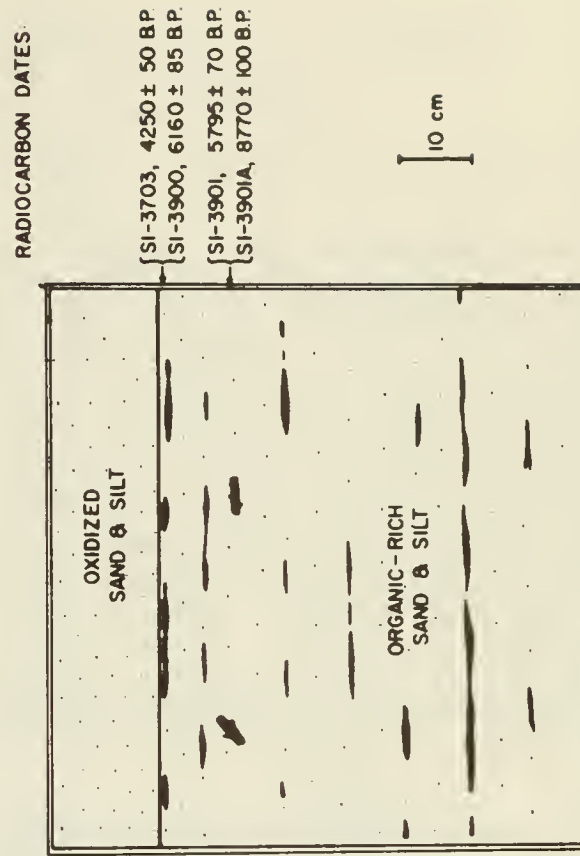


Figure 6. Stop 11: Parent Exposure.

TRIP C-5

SEDIMENTOLOGY OF SILURIAN FLYSCH, ASHLAND SYNCLINORIUM, MAINE

by

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The Silurian of the Ashland Synclinorium offers an opportunity to study turbidity current deposition in a sequence that may well represent continental margin sedimentation following the Taconian Orogeny (Roy, 1978). The uplift of pre-Silurian rocks to the west and northwest of the synclinorium produced emergent terrain with a narrow shelf environment and apparently steep eastward slopes into a basin in which sedimentation was uninterrupted. A 2600 meter (8500 feet) section of Silurian shale, graywacke, and conglomerate was deposited at the base of the slope and in the basin. Basinward from the former land area the deposits become generally finer-grained and more distal in turbidite characteristics. Upward in the Silurian section, similar changes reflect both reduction of land relief and progressive advance of marine conditions onto the detrital source area during the Silurian. These changes produced increasingly more distal conditions everywhere within the basin.

The stratigraphy of the Stockholm region is summarized in Figure 1 and is discussed by Roy and Mencher, (1976) and Roy, Trip B-6, this volume). On this trip we will focus on turbidite deposition within the Jemtland Formation (STOPS 2,4, and 6) which forms part of the upper half of the Silurian section. The lower and more proximal part of the section as seen in the Frenchville and New Sweden Formations will be seen at STOPS 3 and 7; the pre-Silurian stratigraphic setting (Madawaska Lake and Cary's Mills formations) will be examined in STOPS 1,3, and 5. The locations of all stops are given in this volume in Figure 2 of Roy, (Trip B-6).

Post-Taconian Submarine Erosion

We will visit an exposure of the Taconian angular unconformity between the Ordovician Madawaska Lake Formation and the Lower Silurian Frenchville Formation (STOP 3). The details of the exposure are given in the stop description below and the regional aspects of the Ordovician-Silurian transition are given elsewhere in this volume by Roy, (Trip B-6). It is concluded that at least the final pre-Frenchville erosion at this exposure was submarine and it is possible that all of the folding and erosion of the Madawaska Lake beds were accomplished below sea level,

Submarine unconformities with apparently substantial hiatuses in eugeo-synclinal regions are not widely reported as such. Direct and indirect evidence of intraformational submarine erosion with short hiatuses are of course widely observed and reported in both shallow and deep-water sedimentary sections. In the present case the unconformity is angular and is a systemic interface that represents a non-trivial hiatus. The erosion is considered to have occurred on a post-Taconian submarine slope for four reasons:

- 1) To the west and northwest, Lower Silurian rocks are absent,

AGE		COLUMN (Thickness in Meters)	LITHOLOGY AND FEATURES			ENVIRONMENT			NOMENCLATURE					
EARLY DEVONIAN	Emsian	?	760+	Green and red silty shale and slate with interlayered thin beds of very calcareous, laminated fine-grained graywacke and siltstone.			Distal Flysch			Fogelin Hill Formation				
	Siegenian													
	Gedinnian													
SILURIAN	Pridolian	?	760	Thinly interlayered gray shale, non-laminated silty shale, laminated silty shale and fine-grained calcareous graywacke. Medium-grained and coarse-grained graywacke common. Turbidite structures typical of graywacke beds. Abundant graptolites.			Intermediate and Distal Flysch			Jemtland Formation				
	Ludlovian													
	Wenlockian	?	1070-1220	Polymictic conglomerate and lithic graywacke	Quartzose arenite and graywacke, lithic graywacke and slate	Calcareous, phyllitic slate, minor banded Mn-ironstone and graywacke.	Proximal portions of submarine fans	Intermediate portions of submarine fans	Distal portions of submarine fans and basin deposits	Frenchville Fm Conglomerate Member Quartzose Sandstone Member New Sweden Formation				
	Llandoveryian											Upper		
												Middle		
												Lower		
ORDOVICIAN	Ashgillian	?	3800 +1	Unconformity (Locally Angular)	Gray slate thinly interlayered with micritic limestone and minor graywacke ("ribbon rock"); graywacke more abundant in lower part of section.			Submarine Erosion			Taconian Hiatus			
	Caradocian													
								Distal Flysch			Distal Calcareous Flysch		Madavaska Lake Formation Carys Mills Formation	

Figure 1: Stratigraphic section of the Stockholm-Jemtland area (from Roy and Mencher, 1976).

whereas above the unconformity and to the east a thick "deep-water" section representing almost the entire Silurian is present.

- 2) Paleocurrent measurements and provenance of sandstone beds in the Frenchville and Jemtland formations indicate downslope transport of detritus from the west and northwest.
- 3) Frenchville sandstone and conglomerate beds contain brachiopod assemblage composed of representatives of mixed depth communities (Roy, 1973; McKerrow and Ziegler, 1971).
- 4) At least the final erosion of the Madawaska Lake Formation produced large flute molds on the erosional surface which suggests differential scour of semiconsolidated mud by sediment-laden currents.

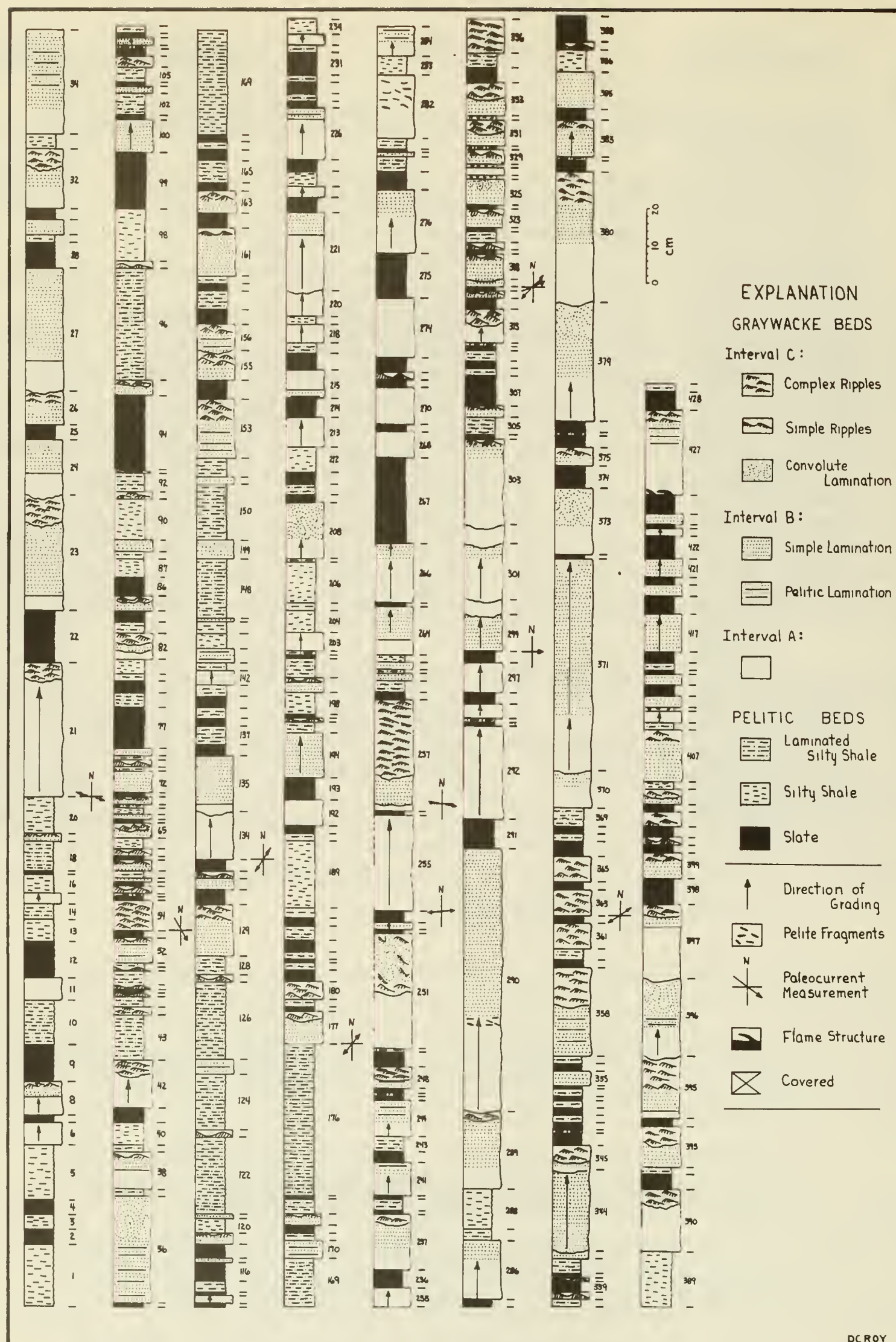


Figure 2; Lithologic and depositional description of the Jemtland section at STOP 4. This is section 2 in Figures 4 and 5.

Early Silurian Sedimentation

Following the Taconian uplift to the west, clastic debris was shed eastward into a broad basin. Along the western margin of the basin, at the position of the Ashland Synclinorium, a thick sequence of sandstone and conglomerate was deposited (Frenchville Formation) which gave way to more shale-and limestone-rich facies out in the basin (New Sweden and Spragueville formations). These Lower Silurian lithofacies are more easily examined in the Ashland-Presque Isle area and are described in more detail elsewhere in this guidebook (Roy, Trip B-6).

Where we are able to see the base of the Frenchville along the western margin the Ashland Synclinorium the formation rests unconformably on the Madawaska Lake Formation as just described in connection with the exposure at STOP 3. The more eastern lithofacies each rest conformably on the "ribbon rock" limestone of the Upper Member of the Carys Mills Formation. It is therefore inferred that the Taconian hiatus decrease eastward across the Ashland Synclinorium to essentially zero and that the Taconian uplift can be fairly well dated by the influx of mud that terminated the more cyclic "ribbon rock" deposition of the Carys Mills. The change from Carys Mills sedimentation to those of the New Sweden and Spragueville formations is pretty well dated as about graptolite zone 19 of the Early Silurian (Pavrides, 1968; Roy and Mencher, 1976). It is likely that the Aroostook-Matapedia Basin in which the Carys Mills was deposited became more shallow as a result of the Taconian uplift that raised land to the west (Roy, 1970a). Subsidence of the basin (and possibly also the western upland) during post-Taconian sedimentation is inferred from the great thickness of the Silurian section in the basin and the fining upward of the section.

Jemtland Sedimentation

The Jemtland Formation is a thin-bedded flysch composed of five principal rock-types and two minor rock-types. The major rock-types are:

- 1) Indistinctly laminated calcareous, platy, silty shale. This rock type contains the most abundant and best preserved graptolites.
- 2) Nonlaminated, calcareous, silty shale. This rock type usually shows an irregular bedding cleavage.
- 3) Dark-gray to black fine-grained thinly cleaved slate. Cleavage in this rock type is usually axial planar to the folds, or nearly so.
- 4) Light-gray, calcareous, micaceous, laminated and cross-laminated, quartzose, siltstone or fine-grained sandstone (graywacke).
- 5) Gray, medium-to coarse-grained, calcareous, feldspathic and lithic graywacke.

These rock types are interlayered on the scale of centimeters and tens of centimeters. It is common for lithic graywacke to grade upward into the micaceous quartzose graywacke in thicker graded beds. The minor rock types are tuff (STOPS 2 and 6) and micritic limestone.

All Jemtland exposures can be described lithologically in terms of the above rocks (usually just the major rock-types) and can be described sedimentologically in terms of a Bouma-type turbidite model. Roy (1970a, 1970b) has shown that the pelitic rock types (types 1,2, and 3 above) do not show any preferred ordering with respect to each other or the graywacke beds and hence must be treated separately in terms of sedimentation. The graywacke beds may be nicely described in terms of Bouma's (1962) A,B, and C intervals. A pelitic bed relates only indirectly to the turbidity current that deposited the underlying graywacke bed as discussed below.

Figure 2 shows a graphical description of the outcrop at STOP 4 (essentially the "type" Jemtland) in terms of Bouma's model as modified by Walker (1965, 1967). This exposure and that at STOP 2 are typical of the graywacke facies of the formation. Figure 3 shows two sections (similarly described) from the more eastern slate facies of the Jemtland; see Figure 9 for the locations of these two eastern sections. From all three of the sections the following generalizations are evident:

1. Amalgamated (composite) turbidites (Walker, 1965) are common in the graywacke facies.
2. Basal "A" intervals are much more common in the graywacke facies. North of the latitude of Washburn there is in fact a fairly regular eastward decrease in the proportion of graywacke beds that begin with basal "A" intervals as shown in Figure 4.
3. The transition from "A" to overlying "B" intervals is usually gradational whereas transitions from "B" intervals to "C" intervals almost always involve an erosional interface. The upward ordering of A,B, and C is statistically favored (Roy, 1970b).
4. Grading is common in "A" intervals and is present but less obvious in many "B" intervals. Graded "A" intervals were produced by deposition under conditions of very rapid fallout. Non-graded "A" intervals may be material transported by grain flows.
5. Both of the silty shale rock types are more abundant in the graywacke facies than the finer-grained dark slate. The dark slate increases in abundance basinward as shown in Figure 5.
6. The most common type of pelite to follow the graywacke beds is the type that is the most abundant in the particular section (Roy, 1970b).

In addition there are features that are not well displayed in the graphical sections of Figures 2 and 3 but are generally observed in outcrop:

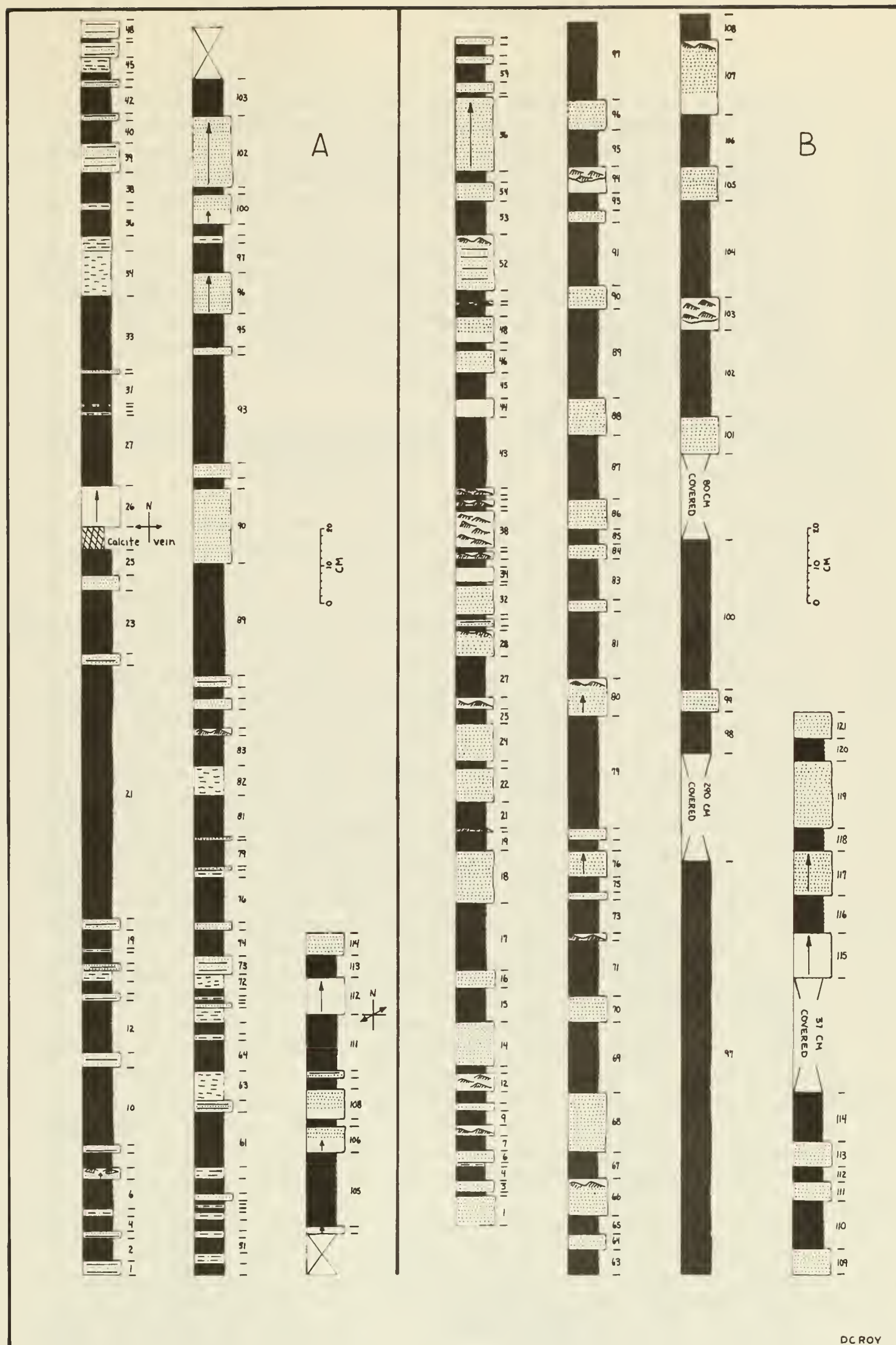


Figure 3: Lithologic and depositional descriptions of outcrops of the slate facies of the Jemtland Formation: A) Outcrop at section 4 of Figures 4 and 5; B) Outcrop at section 5 of Figures 4 and 5 (see also Figure 9).

7. Graywacke beds overlying either of the two types of silty shale described above generally have erosional sole features (flutes, tool marks, ridges and grooves, etc.) whereas soles of graywacke beds overlying the fine-grained pelite show load features (e.g. load casts, flame structures, etc.).
8. The clay/silt matrix of the graywacke beds appears to be primary. This conclusion is based on depositional segregation of the matrix in sedimentary structures and the sharp, little-altered boundaries of feldspar and rock fragments in many of the graywacke beds. Carbonate alteration is more common than clay alteration in the graywacke.
9. Bioturbation is very rare and delicate siltstone laminae within pelite beds are traceable laterally for great distances as is well shown at STOP 2.
10. Laminae within "B" intervals become thinner, more traceable, and more clay/silt rich upward. Pelite laminae are commonly seen in the upper parts of "B" intervals. Weisbrich (1977) and Robert Brown (unpublished data) have produced parallel laminated beds in flume experiments at plane-bed flow velocities in which laminae thickness and, to some extent, clay/silt incorporation are found to depend on the rate of bed aggradation. The experiments were run by allowing sand to fall-out at known rates through a recirculating slurry of fixed lutum (clay plus silt) content. As a rule rapid bed aggradation (high fall-out rate) produces indistinct and thick laminae whereas low bed aggradation (low fall-out rate) produces thin, well defined, and more laterally persistent laminae. Incorporation of lutum in the experimental deposits increases with decreasing bed aggradation rate with marked increases to a few percent at aggradation rates less than about .05 cm/min. So far, pure pelite laminae within experimentally produced laminated fine sand beds have not been formed and their presence in "B" intervals of Jemtland and other turbidites is still a mystery. Variations in laminae thickness and persistence within "B" intervals, however, appears to be useful in qualitatively assessing variations in bed aggradation rates during deposition of these turbidite intervals. The experimental aggradation rates used by Weisbrich and Brown to produce lamination comparable to that seen in the Jemtland ranged from .003 to .5 cm/min; thus the deposition time of "B" intervals of the Jemtland were possibly tens of minutes in duration.
11. Ripple lamination in "C" intervals typically represents climbing ripples with low angles of climb and thus indicate low values of the ratio of bed aggradation-to-ripple migration (Jopling and Walker, 1968; Scheible, 1979). This ratio is given by: $Ar/Mr = \tan Ac$ where Ar is the bed aggradation rate, Mr is the ripple migration rate and Ac is the angle of climb (Scheible, 1979). In fairly well developed climbing ripples, angles of climb of $5 - 10^\circ$ (Ar/Mr of .1 to .2) appear necessary to permit clear discernment of climbing ripple cosets;

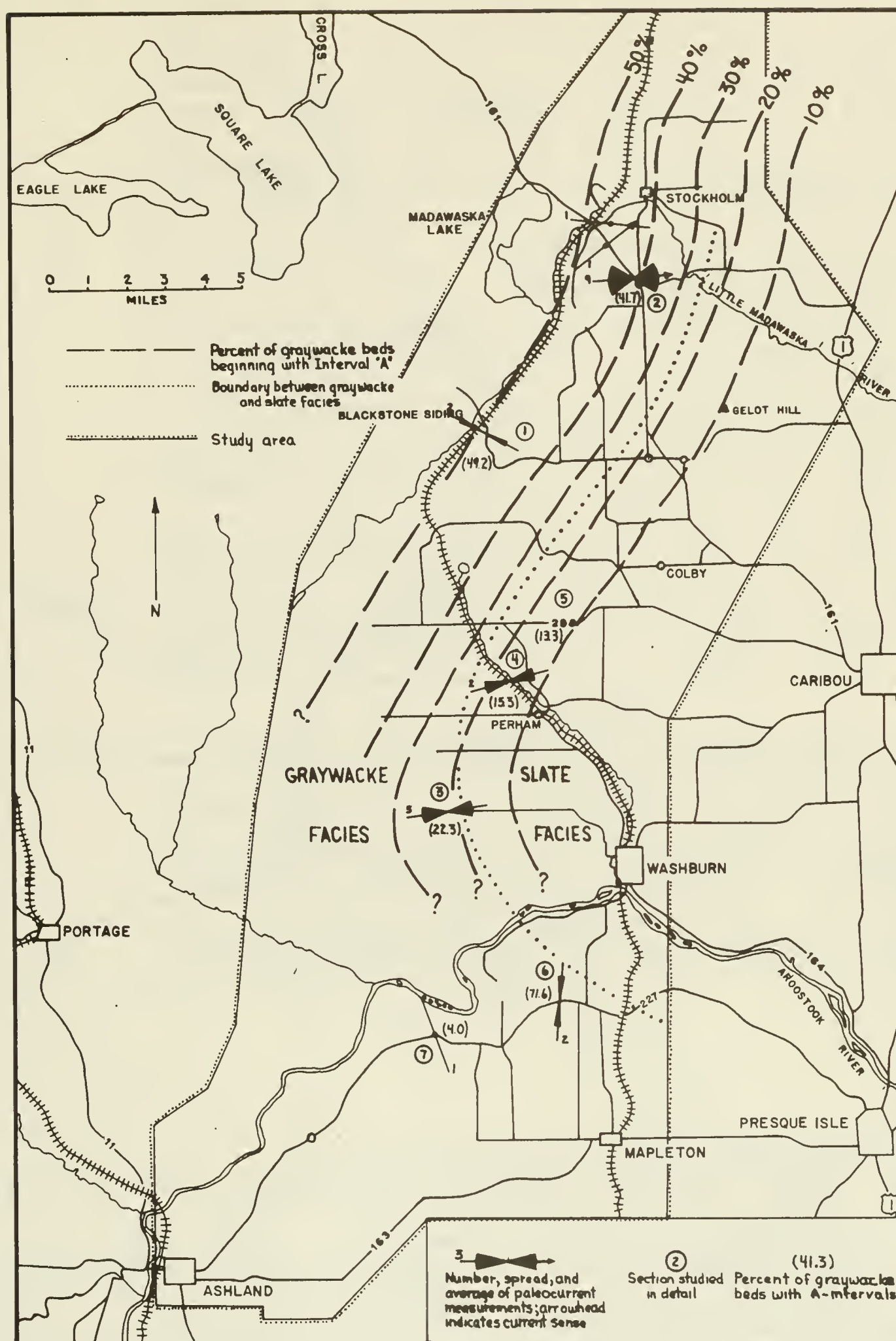


Figure 4: Variation in proportion of graywacke beds in the Jemtland Formation that have basal A-intervals.

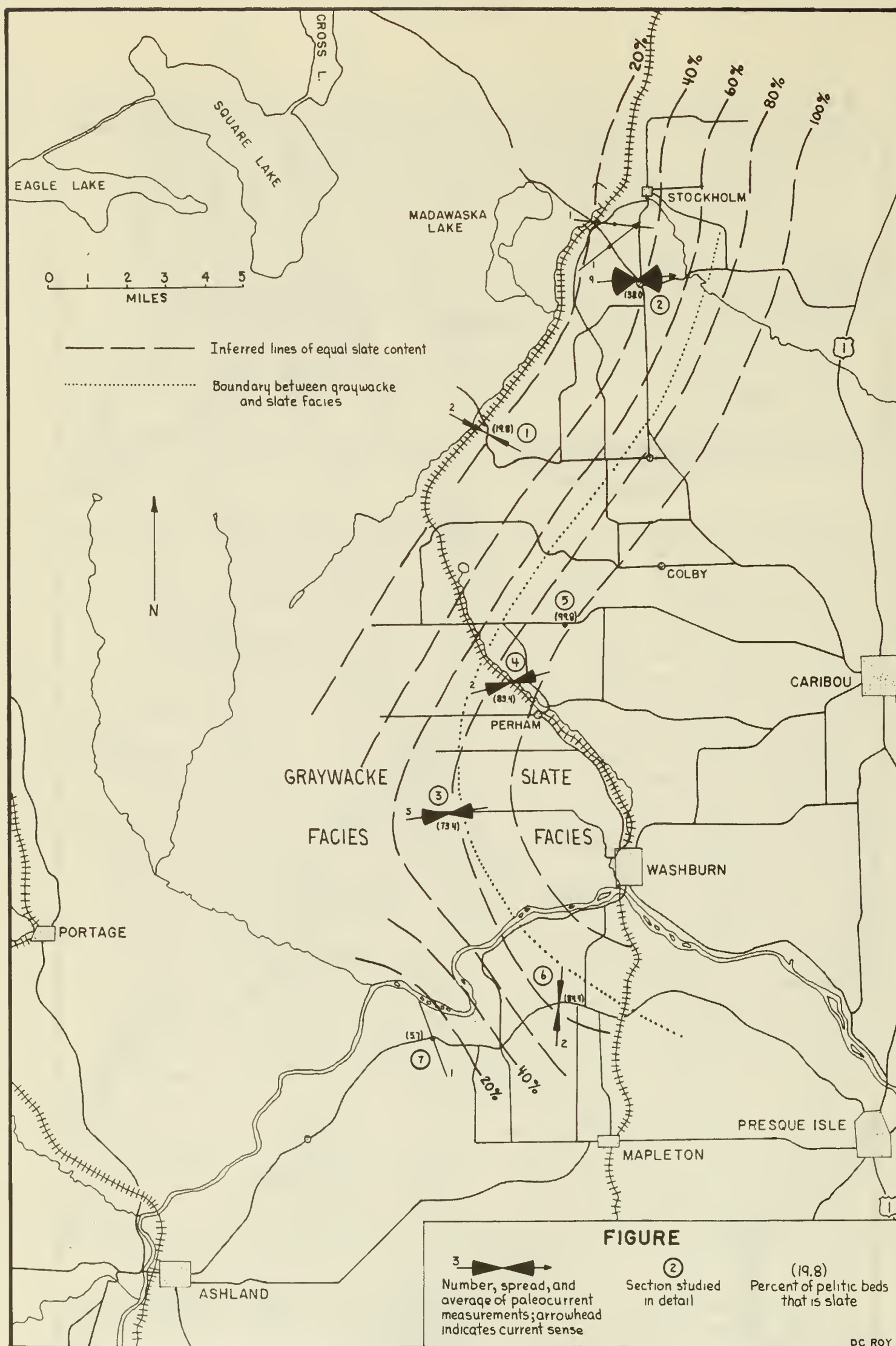


Figure 5: Variation in proportion of fine-grained slate in the Jemmland Formation.

somewhat higher angles of climb are probably necessary in more complicated natural cosets to clearly see the climbing character of the ripples. The generally low Ar/Mr ratios in Jemtland "C" intervals probably is the cause of the erosional surface at the base of most such intervals. Non-erosive transitions from "B" to "C" intervals would require high Ar/Mr ratios or nearly vertical angles of climb to prevent substrate erosion. Scheible's (1979) work suggests that as a practical matter realization of such high Ar/Mr ratios means that flow velocities may have to be below velocities for initiation of motion of the sand grains on a flat bed (to greatly reduce Mr) while appreciable sand of that grainsize is still available in the flow to fall-out. It has been shown that ripples are stable and can migrate at flow velocities less than that required to initiate ripples on a flat bed (Middleton and Southard, 1977). Such low flow velocities, however, may inhibit development of ripple bed forms from the precursor flat bed conditions at the close of "B" interval deposition. This is because the initiation of ripples on a flat-bed requires flow velocities on the order of the flat-bed initial motion velocity unless there is significant perturbation of the flat-bed (Middleton and Southard, 1977). It is also likely that the presence of fall-out from a turbidity current reduces the lower velocity limit of upper flat-bed stability to values well into the ripple stability field as established (for a given grainsize) under conventional non-aggradation experimental conditions. During turbidity current deposition, ripple development is probably suppressed both by fall-out and the continuing reduction of grainsize during sedimentation (Allen, 1970).

12. Pelite clasts are common in the coarser-grained graywacke beds. These, usually tabular, clasts may have maximum dimensions that are on the order of the thickness of the graywacke bed itself. Most of the pelite clasts are fragments of Jemtland shale added to the detritus by submarine erosion upslope from the position of deposition. It is common that the clasts are positioned within the graywacke bed at levels of $1/4$ or $3/4$ of the bed-thickness above the base of the bed. This same localization of pelite clasts is found in Frenchville turbidites and the present writer has no very good explanation for this tendency.

The above features suggest, or are at least consistent with, deposition of the graywacke beds from lutum-rich turbidity currents. The clay and silt of each turbidity current cloud probably remained in suspension for some time following sand deposition. Upward mixing of the warm shelf-water of the turbidity current after reaching the cold bottom water of the basin probably helped to maintain a fairly persistent near-bottom suspension cloud which was re-supplied by subsequent turbidity currents. The unlaminated silty shale and the pelite of the fine-grained slate probably represent deposition from such more or less stagnant suspension clouds. The non-laminated silty shale would represent deposition from the near-source part of the cloud.

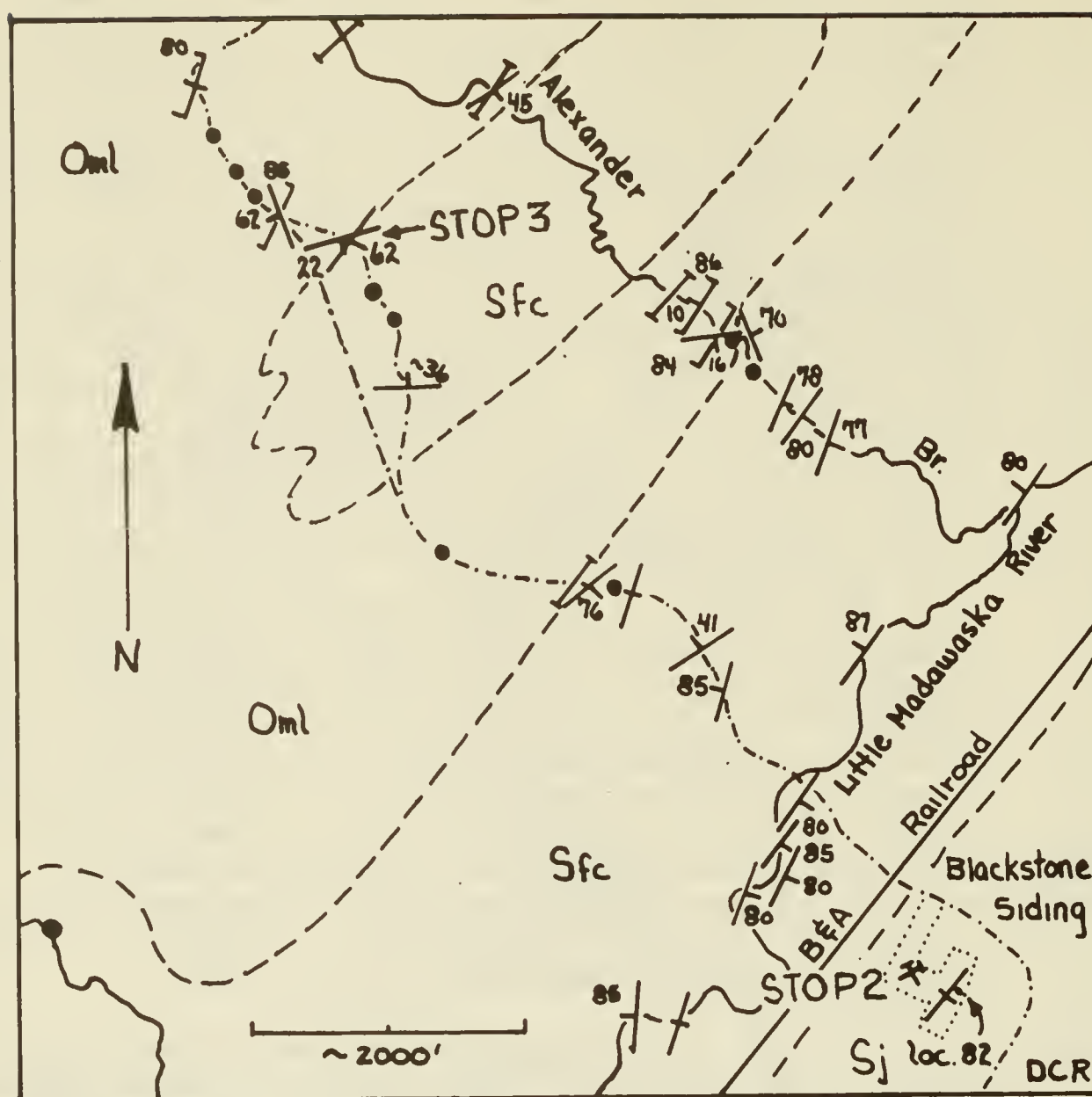


Figure 6: Geologic map of the Blackstone Siding area showing the locations of STOPS 2 and 3. Base from an uncorrected airphoto.

The laminated silty shale is more difficult to explain because the fabric of the shale and the common orientation of graptolites within it suggest current effects during deposition. It is possible that the laminated silty shale beds were deposited from the silt-rich phase of basinal suspension clouds that were moving in response to contour currents, residual turbidity currents, or other currents within the basin.

Post-Jemtland Sedimentation and the Close of the Silurian

Along the axis of the Stockholm Mountain Syncline (Roy, Trip B-6; Figure 2) the Jemtland Formation is overlain by the Fogelin Hill Formation which may span the Siluro-Devonian boundary (Roy and Mencher, 1976; Figure 1). The Fogelin Hill is conspicuous in its content of red and green slate and absence of the coarser-grain graywacke beds found in the Jemtland. The red and green slates suggest important changes in the geochemical environment of the basin following Jemtland deposition. The basin appears to have become more oxidizing and more distal with respect to detrital sources.

The Fogelin Hill Formation appears to span the time interval of the Salinic Disturbance for which there is good evidence both to the east (Presque Isle area) and to the west (Fish River Lake area). Rocks very similar to the Fogelin Hill are present in the Portage area to the west of the synclinorium where they contain Early Devonian graptolites and rest unconformably on Ordovician rocks (see Figure 2 of Roy, Trip B-6; Roy and Mencher, unpublished data). Crinoidal limestone is present as thin beds in both the Portage area and high in the Fogelin Hill section within the synclinorium. The Fogelin Hill Formation is interpreted by Roy, (this volume) to have been deposited in a residual basin of the synclinorium that survived but was restricted by the Salinic "uplifts" on both sides.

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Itinerary

Mileage

Assembly point is the parking lot of Keddy's Motor Inn. Starting time is 8:00 A.M. Head north on U.S.1 through Presque Isle. Mileage starts at the bridge over the Aroostook River north of town.

- 0 Bridge over Aroostook River. U.S.1 from here to Caribou passes through some of the best potato country in Aroostook County. The bedrock supporting the agriculture is composed of the Carys Mills and Spragueville formations.
- 7.75 Bear left off "new U.S.1" onto "old U.S.1" toward downtown Caribou.
- 11.05 Turn left onto Route 161 in Caribou just beyond the downtown shopping mall.
- 11.35 Turn left toward Main Street.
- 11.40 Turn right on Main Street which is also Route 161.
- 12.25 Bear right at junction with Route 228 and continue north on Route 161.
- 16.55 STOP 1. Park on shoulder of highway. Please be careful since the exposure is on the inside of a curve.

This is one of the best exposures of the Lower Member of the Carys Mills as I have mapped it in the Caribou Quadrangle. Graptolites from this exposure have been assigned by Pavlides and others (1961) to zone 13 of the Ordovician (Caradocian). The strata here are about 3500 feet (1200 meters) below the base of the Upper Member of the Carys Mills which is composed of the famed limestone "ribbon rock" (largely of earliest Silurian age) of the Aroostook-Matapedia Belt. Four main rock-types are present here: 1) gray, fine-grained, green-weathering, calcareous slate; 2) light-gray, medium-to-fine grained, micaceous, calcareous, graded, laminated/cross laminated quartzofeldspathic graywacke; 3) light-gray, massive, micaceous, rusty-orange weathering, calcite-veined quartz graywacke in which sedimentary structures are rare; 4) dark-gray, pale-gray weathering, laminated micritic limestone. This pre-Taconian sequence is inferred to have been deposited in an oceanic basin receiving abundant terrigenous detritus from intraoceanic sources which may include volcanic islands to the west and a tectonic island arc in the present Miramichi Anticlinorium (Rast and Stringer, 1975). Terrigenous sources are inferred to wane in importance during deposition of the Upper Member of the Carys Mills where calcareous turbidites dominate the deposit.

Return to cars and continue north on Route 161.

- 19.15 Bear left at Y-intersection. Outcrops of the Late Early Silurian New Sweden Formation are on the left side of the highway. The "Caribou Station" of Boston College's seismic network is located a short distance straight ahead from this intersection (so much for geophysics on this trip).
- 20.00 Turn left (west) at the light in Sweden, Maine (Texaco Station).
- 22.65 Continue straight at the sharp right (north) turn in the paved road. We are entering a major active lumbering area and you are urged to watch out for large trucks; it is further recommended that you let them have the right-of-way.
- 24.15 Turn right (north). Road becomes a bit narrower and rougher here (usually).
- 25.65 Turn left (south) onto the entrance road for the "Blackstone Siding quarry".
- 25.85 Park as best you can.

NOTE: We will spend about 1.5 hours in the Blackstone Siding area (Figure 6) examining STOPS 2 and 3. The outcrop at STOP 3 is small (!) and is best examined by small groups so it is suggested that cars go to it one or two at a time (following directions below) during our stay here. Those more interested in studying Jemtland turbidites may simply stay at STOP 2.

STOP 2. The Blackstone Siding quarry provides the best exposure of the Jemtland Formation in the region. It displays the most proximal phase of the formation and provides slightly weathered pavement surfaces that permit careful study of turbidite features. The strata

strike about N35°E, are essentially vertical, and top to the southeast. No reversals of facing have been found. In the southeast corner of the pit are exposures of basal beds of the Aquagene Tuff Member of the formation which can be traced discontinuously to the northeast for a distance of about nine miles (14 km). Graptolites are ubiquitous in this quarry and some fine specimens are possible. The strata here are of Early Ludlovian age (graptolite zones 33-34 of the Silurian).

Return to cars and head out to the main road.

- 26.05 Main lumber road. Turn left (west).
- 26.15 Blackstone Siding (International Paper). In this vicinity we cross the Jemtland-Frenchville contact that is not observed here but is seen to be gradational in river exposures along strike to the northeast.
- 26.30 Little Madawaska River.
- 26.3 Road to right (north); stay on main road.
- 26.8 Turn right onto a segment of the old haul road that is now been bypassed.
- 27.3 STOP 3. In this small outcrop (some have called it an "incrop") the Taconian unconformity is exposed between the Madawaska Lake Formation (Ordovician) and the Frenchville Formation (Figure 6). The contact is interesting because it is angular, there are large flute casts on the sole of the basal sandstone bed of the Frenchville, and step-lineations on the sole of the sandstone preserve the intersection of Madawaska Lake bedding and the erosion surface. The relationships of these features are shown in Figure 7A. The flute casts are assymmetric with the steep slope of the flute located on the down-dip (pre-Frenchville) side of the flute and the "floor" of the flute mold was essentially parallel to the bedding in the underlying Madawaska Lake Formation (Figure 7B). Taken together, these features suggest Taconian folding (submarine?) of semiconsolidated Madawaska Lake beds followed by submarine erosion by a turbidity current and finally deposition of the pebbly lithic graywacke. The material of the graywacke bed may have been transported by grainflow or other mechanisms, arriving at the bottom position (canyon floor?) represented by the outcrop after the turbidity current had largely passed.

Return to cars and continue northwest along the old stretch of the lumber road.

- 27.40 Main road. Make sharp left to head back (eastward) on the main road.
- 27.5 As you go up the hill the Madawaska Lake-Frenchville contact is again crossed (we are only a short distance along strike from STOP 3).
- 28.30 Little Madawaska River.
- 28.45 Blackstone Siding.

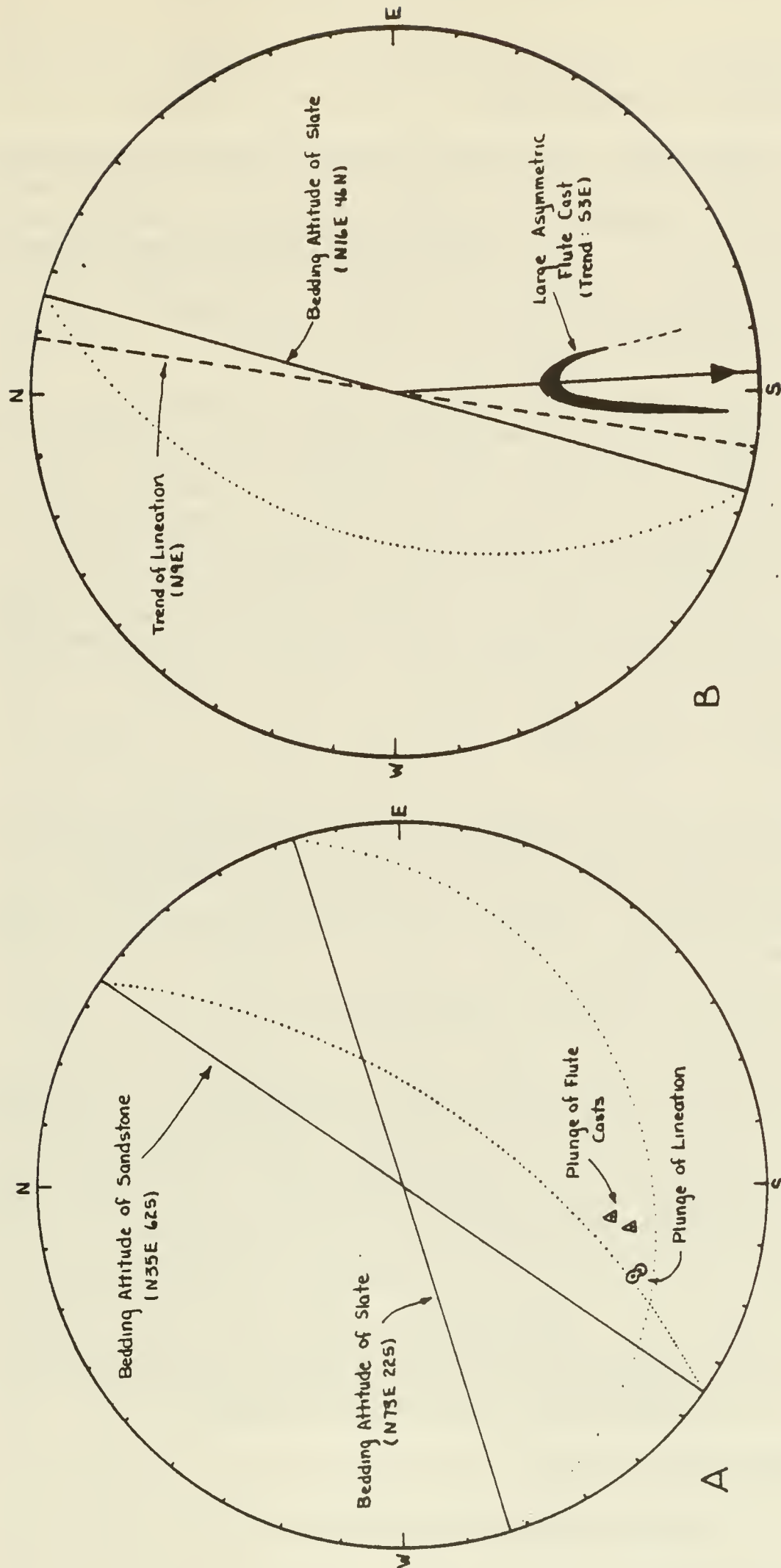


Figure 7: A- Present relationships of bedding step-lineation, and flute-cast plunge at STOP 3.
 B- Pre-Frenchville orientations of Madawaska Lake bedding and flute casts showing the relationship of the cast asymmetry to bedding in the substrate.

- 28.55 Entrance to Blackstone Siding Quarry. Continue east.
- 30.05 Turn left (east).
- 31.55 Continue straight on paved road.
- 34.20 Intersection with Route 161 (Texico Station). Turn left (north).
- 38.80 STOP 4. Park as directed. The exposure is on the west side of Route 161 in a slight curve. Care must be exercised in both parking and crossing the highway. This will probably be a good place for a LUNCH STOP. A small store (if open on Sunday) in Jemtland (near the stop) may provide provisions.

This is the "type exposure" of the Jemtland Formation. The formation here is similar to that seen at STOP 2 but is more slate-rich (less silty shale) and is inferred to be slightly more distal. The outcrop is near the middle of the Jemtland and is on the southeastern flank of the Stockholm Mountain Syncline (see Figure 2 of Roy, Trip B-6, this volume). The proportion of graywacke (58 percent) is similar to that in the sequence at STOP 2. Figure 2 shows the details of the stratigraphic section here. Common sole features readily permit paleocurrent measurements which show a west-to-east flow of the turbidity currents (Figures 2, 4, and 5).

Continue north on Route 161.

- 39.50 Approximately here we cross the axis of the Stockholm Mountain Syncline. Youngest unit is the poorly exposed Fogelin Hill Formation.
- 40.75 Little Madawaska River. Frenchville Formation is exposed in the river upstream from the bridge.
- 43.30 STOP 5. Park on the shoulder of the highway.

This is one of the better exposures of the Madawaska Lake Formation (Ordovician) in the "type area" of the formation. In this exposure the rocks are typical and consist of predominant olive-green, fracture cleaved slate with lesser thin beds and laminae of calcareous quartzofeldspathic siltstone or fine-grained sandstone. Here also are thin beds of rusty-weathering pyritiferous micritic limestone. One exposure here also shows red slate which is a minor but conspicuous rock type in the formation near the unconformity with the overlying Frenchville Formation in this area. The Madawaska Lake is the temporal equivalent of the Winterville Formation (volcanic rich) to the west. More westerly in the Madawaska Lake, lithic graywacke increases in abundance presumably reflecting proximity to volcanic islands.

Continue north on Route 161.

- 43.45 Turn around in the Rest Area and head south.
- 46.15 Little Madawaska River.

- 46.20 Turn left (east) on paved road.
- 47.70 Turn left (north) at Stockholm Post Office. Continue north across the Little Madawaska River and R.R. tracks, by the store, and up the hill.
- 48.60 Entrance to Stockholm Town Dump on left. Enter and drive to refuse area and park. Town permission is required to enter this scenic area. Walk southwest along the strike of the lithic tuff horizon that is more or less exposed in the dump.

STOP 6. Aquagene Tuff Member (provisional name) of the Jemtland Formation. An almost complete section of the tuff sequence is available in these brush covered exposures and is depicted in Figure 8. The member (here 63 feet thick) is stratigraphically in the middle of the formation on the northwest flank of the Stockholm Mountain Syncline as shown in Figure 2 of Roy, (Trip B-6, this volume). The member is thinner and more discontinuous on the southeast flank of the fold. Fossil localities stratigraphically below, above, and within the member in the immediate vicinity date the tuff as Ludlovian of the Late Silurian. Four broad tuffaceous rock types comprise this section. Type I tuff, the most abundant, is light gray-green, chalk-white weathering, and variably lithic. In section devitrified shards and pumice fragments are clearly visible. Plagioclase, quartz, and chlorite are dominant mineral phases. Lithic tuff, designated as Type II, is composed of sand-to pebble-size fragments of intermediate and mafic volcanic rocks, plagioclase, quartz, myrmekitic quartz-feldspar, and detrital carbonate (including fossil fragments) set in a matrix of devitrified shards and pumice fragments. Type II tuff usually transitions upward into Type I in apparently graded sedimentation units. Type III tuff is a chlorite-rich rock that differs from Type I primarily in the presence of large chlorite "patches" that are flattened parallel to a bedding parting. Upward gradation from Type I to Type III tuff is observed in three beds (Figure 8). Type IV tuff is fine-grained, light gray-green in color, cherty in appearance, and occurs in thin beds less than 6 inches thick that commonly show faint lamination. This type forms a 7-foot interval in the middle of the section. In thin section tuff of Type IV shows devitrified shards and pumice fragments in a dominant matrix of microcrystalline quartz-chlorite-plagioclase. The sequence of tuff here is thought to have been deposited in deep water and probably consists largely of resedimented tuffaceous material. The ash was most likely derived from the west where contemporaneous volcanic rocks interlayered with shallow-water sediments are present.

Return to cars and return to main road.

- 48.9 Main road at entrance to the dump. Turn right.
- 49.8 Intersection at the Stockholm Post Office. Turn right (west).
- 51.3 Intersection with Route 161. Turn left (south).
- 53.2 Outcrop of STOP 4 on the right.
- 57.8 Intersection at light (Texaco Station). Continue south.

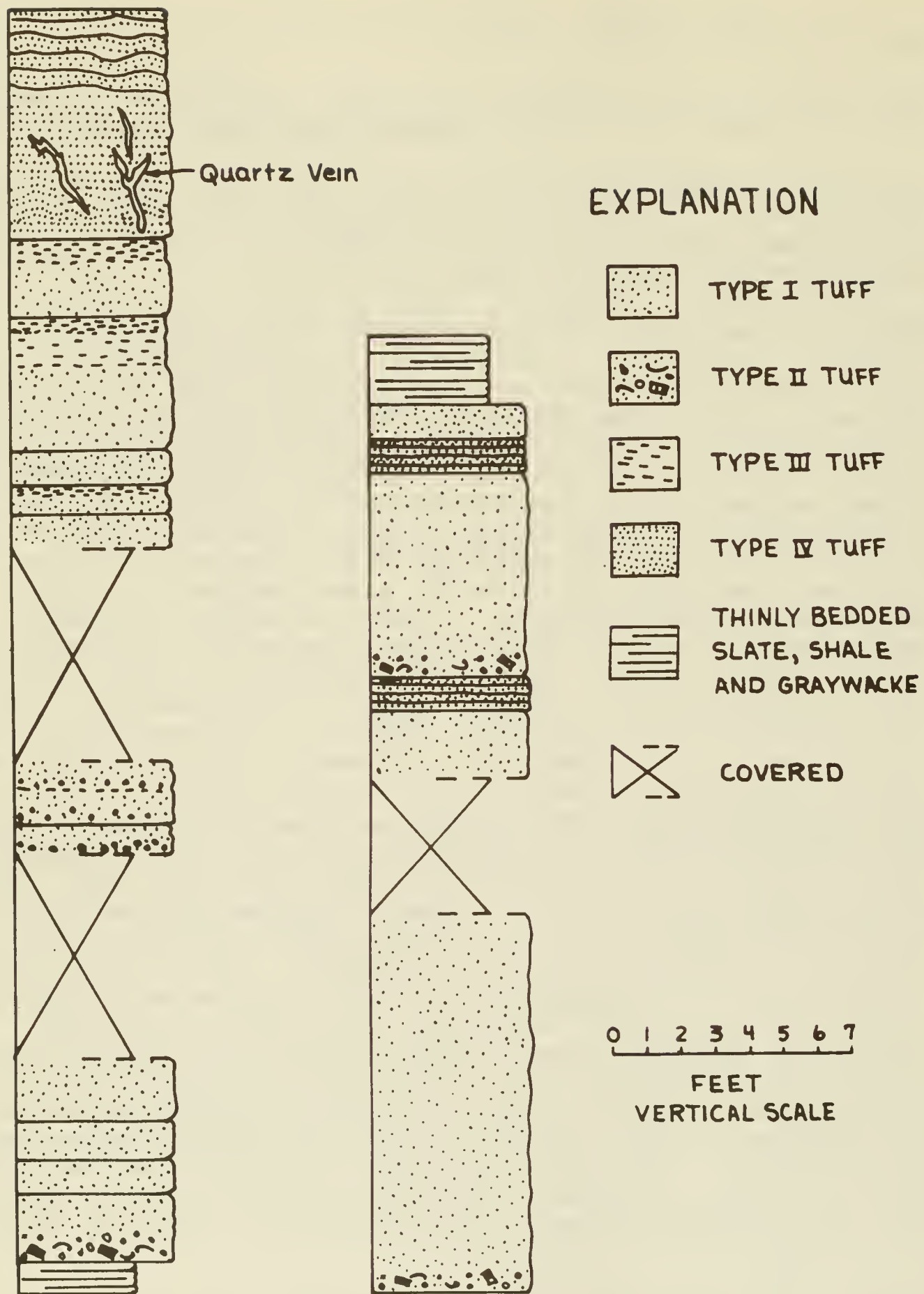


Figure 8: Stratigraphic section of the Aquagene Tuff Member of the Jemtland Formation at STOP 6. The types of tuff are described in the stop description.

- 60.4 Turn right (southwest) onto gravel road.
- 61.05 Turn right (west) onto paved road.
- 62.25 Railroad crossing in Colby.
- 63.3 Turn a hard left (southeast) onto gravel road at "T" intersection.
- 64.4 Turn right (west) onto paved road (Route 228).
- 65.35 Turn left (south) onto gravel road.
- 67.75 Turn right (west) onto paved road.
- 69.15 Junction with Route 228 in Perham. Turn left (south).
- 69.35 Main intersection in Perham. Turn left (southeast) and park in church parking lot (if vacant).

STOP 7. In the village of Perham there are at least three mappable horizons of manganiferous ironstone within the New Sweden Formation. These horizons and others in the Perham area are shown in Figure 9. Here the ironstone is interlayered with gray, calcareous, finely cleaved, laminated, phyllitic slate with lesser micritic limestone beds. Facing here is to the southeast. The ironstone is thinly laminated and generally quite calcareous. The details of laminae mineralogy have not been determined as yet, but hematite-rich (including specularite), carbonate-rich, and manganese-rich varieties seem most prominent. The manganese-rich laminae are reported to contain predominantly Braunite ($3\text{Mn}_2\text{O}_3 \cdot \text{MnSiO}_3$) by Pavlides (1962); Braunite is also found in small ovoid masses as well as in laminae. The origin of the laminated ironstones is not well understood. They appear to be primary sedimentary deposits which have been altered by diagenesis and low-grade metamorphism. They occur at various stratigraphic levels within the New Sweden and are rarely more than a few feet thick and a few hundred feet in strike-dimension. They are commonly interrupted by "barren" slate intervals. The New Sweden is a deep-water formation, approximately 4000 feet thick, that is distal to the turbidite-rich phases of the Frenchville Formation. The iron and manganese have a logical source in the volcanic rocks of the Ordovician Winterville Formation which formed the large emergent terrain (Taconia) a "short" distance to the west. I rather suspect that the laminated ironstones are explicable in terms of turbidity current transport of shallow, warm, shelf water (plus sediment), with the iron and manganese in both particulate and soluble forms, into a deep-water basin. The details await a better understanding of the laminae mineralogy and trace-element chemistry.

The road log ends with STOP 7. Participants may return to Presque Isle by continuing south on Route 228 to Washburn and taking Route 164 to Presque Isle. Route 164 intersects U.S.1 just north of Presque Isle. Presque Isle is about 18 miles from here. Those wishing to go to Caribou simply return to the intersection at mileage 69.15 and turn right (east); you will pass through Carson and join Route 164 in Jacobs and thence to Caribou. Caribou is about 12 miles from here.

Figure 9: Geologic map of the Perham area showing the locations of STOP 7 and sections A and B of Figure 3. Lines of x's indicate traces of manganese horizons; dotted lines show outlines of cleared areas. Base is an uncorrected airphoto.

TRIP C-6

WISCONSINAN GLACIATION OF EASTERN AROOSTOOK COUNTY, MAINE

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Introduction

The purpose of this trip is to examine the Van Buren and Mars Hill Drifts which constitute the surface glacial deposits of eastern Aroostook County and relate them to models of late Wisconsinan glaciation. In addition, moraines extending from St. Francis through Caribou to Fort Fairfield, and moraine complexes at Mars Hill and Lower Macwahoc will be examined (Fig. 1). We will use the field evidence that relates to the problem of the origin of the drifts and the moraines to suggest deglaciation models.

Maps: Mattawamkeag, Wytovitlock, Sherman, Mars Hill, Caribou, Presque Isle, 15 minute series.

Late Wisconsinan Glaciation

The late Wisconsinan ice advance had reached its maximum extent onto the continental shelf and had begun to retreat by 17,000 years BP. (Connally and Sirkin, 1973). Between 13,500 and 12,500 years BP. the active ice margin had retreated from the present coastal position leaving a belt of submarine moraines (Borns, 1966, 1967, 1973; Stuiver and Borns, 1967; Stuiver and Borns, unpub. data). The general recession that produced the coastal moraine complex was interrupted by a major readvance in eastern Maine that culminated in the sea at Pineo Ridge approximately 12,700 years ago (Borns, 1967).

While the receding ice margin was depositing submarine moraines along the present coast of Maine, the Champlain Sea transgressed up the St. Lawrence Lowland from the Gulf of St. Lawrence reaching the vicinity of Ottawa, Ontario, by 12,800 years BP. (Richard, 1978). This Champlain Sea penetration cleaved the Laurentide Ice Sheet along the trend of the St. Lawrence Lowland. The cleaved ice sheet to the south of the Champlain Sea formed a residual ice cap over southern Quebec, Maine, and New Bruns-

*Work supported by the Maine Geological Survey

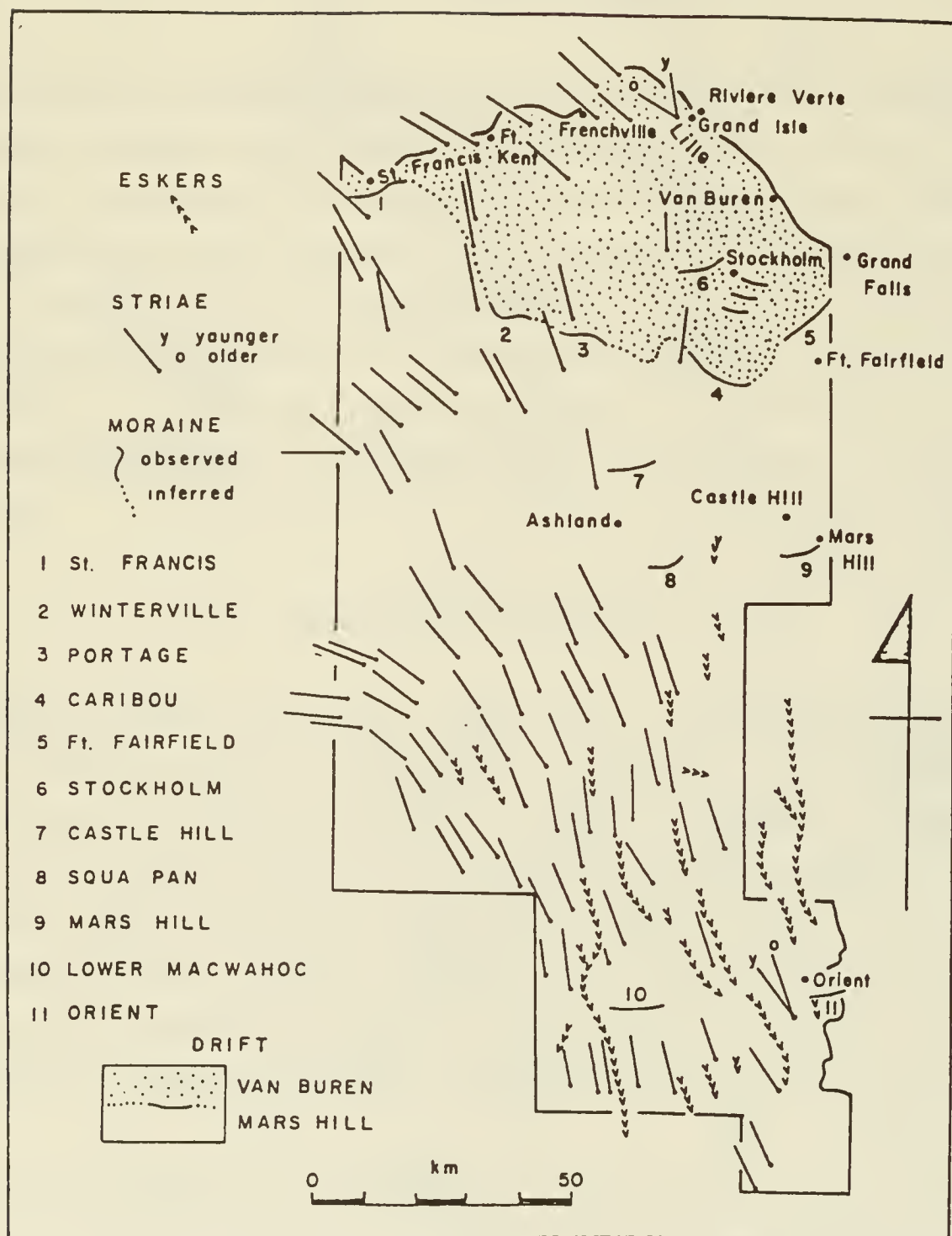


Fig. 1. Generalized surficial geology of Aroostook County.

wick after 12,800 years BP. (Borns and Hughes, 1977). This represents the minimum date at which the Laurentide ice in Maine became detached from the Laurentide ice north of the St. Lawrence River.

Deglaciation of the newly separated ice cap proceeded through downwasting and recession. Residual ice cap conditions in northern Maine, and probably adjacent New Brunswick, lead to the formation of occasional moraine complexes, outwash deposits, and eskers. Peat overlying till along the Green River in New Brunswick suggests that northern Maine was deglaciated sometime before 10,200 years BP. (Kite, 1979).

Glacial Drifts

Two distinct late Wisconsinan glacial drifts mantle eastern Aroostook County (Genesand Newman, 1978, 1979). The Van Buren Drift, covering the very northern portion of the county (Fig. 1), includes a compact, silty, buff to dark brown till containing 3-5% clasts of Precambrian granite gneiss. The Mars Hill Drift covering southern Aroostook County and beyond, includes a loose, sandy, brown till. Clasts in this drift include DeBoullie Granodiorite, Chapman Sandstone, and Mapleton Sandstone - lithologies characteristic of regions immediately to the northwest. Granite gneiss inclusions have yet to be observed in this till by the authors.

Directional indicators point to emplacement of both tills by ice moving in a southeasterly direction but no areal or stratigraphic contact between these tills have been found.

Moraines

The moraines of Aroostook County are generally low, narrow, undulating structures that are indistinguishable at times from bedrock controlled topography and irregularities of dense forest canopy. The Lower Macwahoc Moraine was discovered only after a clear cutting operation permitted a view of the surface morphology. The moraine complex at Caribou closely approximates the southern margin of the Van Buren Drift and consists of Van Buren Till.

The orientation and position of the moraines, meltwater channels, outwash, and eskers indicate that glacial recession was accomplished by

downwasting and recession from central Maine to northern Aroostook County.

Glaciation Models

In light of the available data, two distinct models for the late Wisconsinan history of northern Maine are proposed to explain the observed relationship between the Van Buren and Mars Hill Drifts:

- a) Separate Ice Cap Model: A separate ice cap developed in central Maine during mid Wisconsinan time which eventually expanded and interacted with the advancing Laurentide Ice Sheet from north of the present St. Lawrence Valley. Eventually they coalesced and the Maine ice cap was displaced to the southeast. The Van Buren Till was lodged beneath the Laurentide ice and the Mars Hill Till was lodged beneath the Maine ice.
- b) Frozen Bed - Melted Bed Model: The late Wisconsinan Laurentide ice cap extended across Maine and onto the continental shelf. Both the Van Buren and Mars Hill Till were lodged by this ice advance. The presence of tills containing Canadian Shield erratics (Van Buren) or local rock types (Mars Hill) was a function of the thermal regime that existed at the ice-bedrock interface.

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Itinerary

Mileage

- | | |
|------|---|
| 0 | Assembly point is parking lot at Keddys Motor Inn, Presque Isle. Starting time 8:00 A.M. Go north on route 1. Follow signs to Caribou. |
| 9.6 | Take left on 161 north and continue on route 161. |
| 18.9 | Take right off 161 and head up hill. |
| 19.6 | <u>Stop 1.</u> Caribou Moraine. Here, gravel pit operations have exposed a section 10-12 meters high permitting us to examine the morphology and geographic relationship of the moraine to bedrock hills. Note the approximately 3-5% occurrence of granite gneiss clasts |

in the exposure. These erratics are derived from the Laurentide region of Quebec. South of this locality we have yet to find granite gneiss in tills. This moraine, is part of a discontinuous moraine complex which extends from St. Francis through Caribou and Fort Fairfield, and into New Brunswick where it is correlated with the Grand Falls Moraine.

Return to cars promptly and return the same way as entered.

20.2 At foot of hill take left onto route 161 south.

22.9 Turn left into Caribou Country Club off 161 south.

Stop 2. Caribou Country Club. Drive to the top of the ridge by the club. From this vantage point we have a good view of the morphology of the moraine. Low undulating ridges extend northerly toward the cut seen on stop 1. Although the ridge has been smoothed it is a natural feature in excess of 12 meters thickness. Cuts behind the locality expose granite gneiss inclusions within the till comprising the moraine.

Return to cars.

23.4 Return to 161 south and turn left. The bedrock ridges are perpendicular to the drift ridges in this region. From this point southward no granite gneiss inclusions have been found in Aroostook County tills.

26.3 Take right on route 161. Head toward Presque Isle.

27.0 Take left onto route 1.

27.2 Take right to Presque Isle on route 1. We will follow the beautiful Aroostook River Valley. Once we leave Caribou and head south toward Presque Isle the topography is smooth and regular reflecting bedrock control. Note the clumps of boulders in the fields. These are purposely left by the farmers so that they can identify or locate areas where bedrock is close to the surface and thus avoid damage to their plows.

38.3 Cross the Aroostook River.

39.8 Northland Hotel on left.

40.6 University of Maine at Presque Isle on right.

48.7 Turn left off route 1 south toward Westfield. We are observing a change from bedrock controlled topography to hummocky constructional topography. This is the beginning of the Mars Hill Moraine complex.

50.3 Note the cut into the pit on the right as we swing left.

- 50.5 Turn right. We have entered the Mars Hill Moraine complex.
- 50.8 Note Carry's Mills bedrock exposure on right. This is one of the few bedrock exposures in the area. Mars Hill is seen directly ahead.
- 51.9 Turn right at the "T". All topography in this region is constructional. As we continue south note the increased frequency of moraines being cut by rills and meltwater stream channels. On the right is Green Mountain Ridge. Mars Hill and Green Mountain Ridge funneled late Wisconsinan ice thus constricting its flow resulting in the formation of the Mars Hill Moraine complex.
- 55.3 To the right is a former glacial meltwater channel.
- 55.7 Take left on secondary road by large partially dissected moraine ridge.
- 55.8 Cross route 1A onto road directly across highway.
- 55.9 Stop 3. Mars Hill Moraine Complex. Park cars along the road. The thickness of the moraine ridges varies from 0-30 meters. The complex is an irregular, hummocky sheet consisting of outwash, kames, kettles, and smaller till hummocks. All of these forms have been dissected by outwash streams that flowed directly from a receding ice margin. Clasts in this drift include the DeBoullie Granodiorite, Chapman Sandstone, and Mapleton Sandstone—all local lithologies. Clasts of Mars Hill Conglomerate within the till appear only at the southern end of the till complex and indicates lodgement by till with a southern flow. An extensive moraine complex characterized by distant moraine ridges rises above the general level of the drift surface. At Mars Hill, Squa Pan Lake, Macwahoc, and Orient these ridges take the form of frontal moraines, some of which are associated with outwash. The Mars Hill Moraine complex is the most striking. It extends from Mars Hill (495m elevation) westward to Green Mountain Ridge (360m elevation) and thence northward approximately 16-24 kilometers to Easton and Phair.
- Return to cars and go straight ahead.
- 56.5 Take right fork. Cemetery on left. We crossed a large former meltwater channel just prior to the road intersection.
- 57.3 T-junction. Take right fork. Note possible lateral moraine flanking Mars Hill.
- 60.2 Left off main road.
- 60.6 Take left onto L.L. Boyd and Son's Farm.

Stop 4. Boyd Pit.

This exposure in the Mars Hill Moraine is near the distal terminus of the moraine. To the south of this ridge the region is characterized by outwash (stratified drift). The Mars Hill Conglomerate derived from Mars Hill comprises many of the clasts in this pit.

Turn right out of the pit and farm. The hill to the left is in New Brunswick.

- 61.2 Turn right at junction. Travel north along Mars Hill noting Green Mountain Ridge to the left. As you approach Mars Hill bedrock approaches the surface.
- 64.0 Turn left and follow dirt road.
- 64.9 Continue on asphalt road.
- 65.6 Turn left on route 1A and head for the town of Mars Hill. Note that the topography begins to level out toward the south.
- 66.6 1A joins route 1 south. Continue down route 1 to route 95. You are now entering a pitted outwash plain. From Mars Hill to route 95 is a dangerous road with frequent accidents involving trucks. Be aware.
- 93.7 Take right onto 95 south at Houlton.
- 132.0 Take Sherman exit off 95 south.
- 132.3 Turn left on route 158 toward Sherman Mills. Striation locality.
- 133.8 Bear right beyond Gulf station. Stay on route 158 heading toward Macwahoc.
- 136.6 Take right on route 2.
- 142.7 Take left by old shingle shack off route 2. This dirt logging road is designated 06-20-10.
- 145.6 Take right onto dirt road 6.
- 148.4 Bridge over the Lower Macwahoc River. Park the cars along the road. Be certain that your car is well off the road as this is a main logging road with the trucks really rolling.

Stop 5. The Lower Macwahoc Till. This is a weathered, silty till very similar in texture to the St. Francis Till, which underlies Van Buren Till, at localities along the St. John River. This stop is approximately 2.7 miles southeast of the Lower Macwahoc Moraine, which will be our next locality.

Return to cars and continue on road 6.

- 148.9 Junction with road 6 and 7. Take left onto 7.

- 150.0 Junction with road 7 and 8. Take left onto 8.
- 150.3 Junction with road 8 and 8.1. Continue on road 8.
- 150.8 Junction with road 8 and 8.2. Take a left onto road 8.2.
- 151.2 Stop 6. Lower Macwahoc Moraine. This moraine was exposed during salvage operations of a spruce budworm infested forest. The moraine has been traced for approximately two miles before becoming obscured by dense vegetation. The fabric of the imbreccated clasts indicate\$ lodgement by ice flowing south.
- Return to cars and return to route 2.
- 160.2 Take left onto to route 2 and head toward Macwahoc.
- 167.6 Molunkus Stream picnic area. The road now follows an esker dipping on and off for a short distance.
- 169.9 At Molunkus take right on route 2.
- 178.7 Mattawamkeag. Take right on route 157 and head toward Millinocket.
- 186.9 Crossing the Salmon River.
- 187.3 Take left into cut of the Penobscot Esker.
- Stop 7. Penobscot Esker gravel pit. This locality is at the approximate latitude where the esker systems of northern Maine begin. To the north, only small and insignificant esker forms are encountered.
- Return to cars and head back to road.
- 188.7 Take left onto route 151 upon leaving the pit.
- 190.4 Junction with route 95 Interstate.

End of Trip

STRUCTURE AND SEDIMENTOLOGY OF SILURO-DEVONIAN
BETWEEN EDMUNDSTON AND GRAND FALLS, NEW BRUNSWICK

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Introduction

The Ordovician, Silurian and Lower Devonian sediments of the Edmundston - Grand Falls area in northwestern New Brunswick are exposed in Acadian (Devonian) anticlinoria and synclinoria (Fig. 1) within which subsidiary major and mesoscopic folds are present (Hamilton-Smith, 1970; St. Peter, 1977). The anticlinorium near Grand Falls (Fig. 1) is part of the Aroostook-Matapedia anticlinorium which extends from Maine across New Brunswick to the eastern end of the Gaspé peninsula in Quebec (Rodgers, 1970; Williams, 1978). The lithologies of the folded formations and their stratigraphic correlation within the area are indicated in Table 1. The sediments are predominantly fine-grained greywackes and constitute the Aroostook-Matapedia belt.

Tight upright folds and an associated steep to vertical cleavage are characteristic structural features of the belt. As noted by Rodgers (1970, p. 131), the folding must represent a very great shortening of the original basin across the strike such that the present width of the belt is only a relatively small fraction of its original width. No sign of the floor on which the Ordovician to Lower Devonian sediments were laid down is visible anywhere in the belt, and Rodgers (1970) suggested that if the floor was shortened with the sedimentary cover it must have been non-sialic and presumably was disposed of downward [by subduction]. Ordovician ocean crust of probable Lower Ordovician age has been recognized, however, as basement to the folded and cleaved Silurian-Lower Devonian strata of the Chaleur Bay synclinorium southeast of the Aroostook-Matapedia anticlinorium in northern New Brunswick (Stringer, 1975; Rast *et al.*, 1976; Pajari *et al.*, 1977; Rast and Stringer, 1980). Shortening of the cover rocks by decollement on the underlying basement has been proposed (Rodgers, 1970, p. 132; Stringer, 1975).

The folding and cleavage of Silurian-Lower Devonian rocks in the Appalachian/Caledonian orogen have been attributed to late Silurian and Devonian deformation resulting from continental collision during closure of the Proto-Atlantic (Iapetus) Ocean (Dewey and Kidd, 1974), and in southern Scotland the deformation has been related directly to subduction during the Silurian (Leggett *et al.*, 1979; Stringer and Treagus, 1980) along the northwest margin of the Iapetus Ocean (Cocks *et al.*, 1980). In the Canadian Appalachians, however, Williams (1979) has argued that the Iapetus Ocean closed during the Taconian (Ordovician) orogeny and that the Acadian orogeny was intraplate, subsequent to the closure of the Iapetus Ocean. In New England, Robinson and Hall (1980) have suggested that the Acadian orogeny was a result of Siluro-Devonian convergence between adjoining continental plates, because closure of an ocean cannot be proven and it is possible that the tectonic features formed through development of a west-

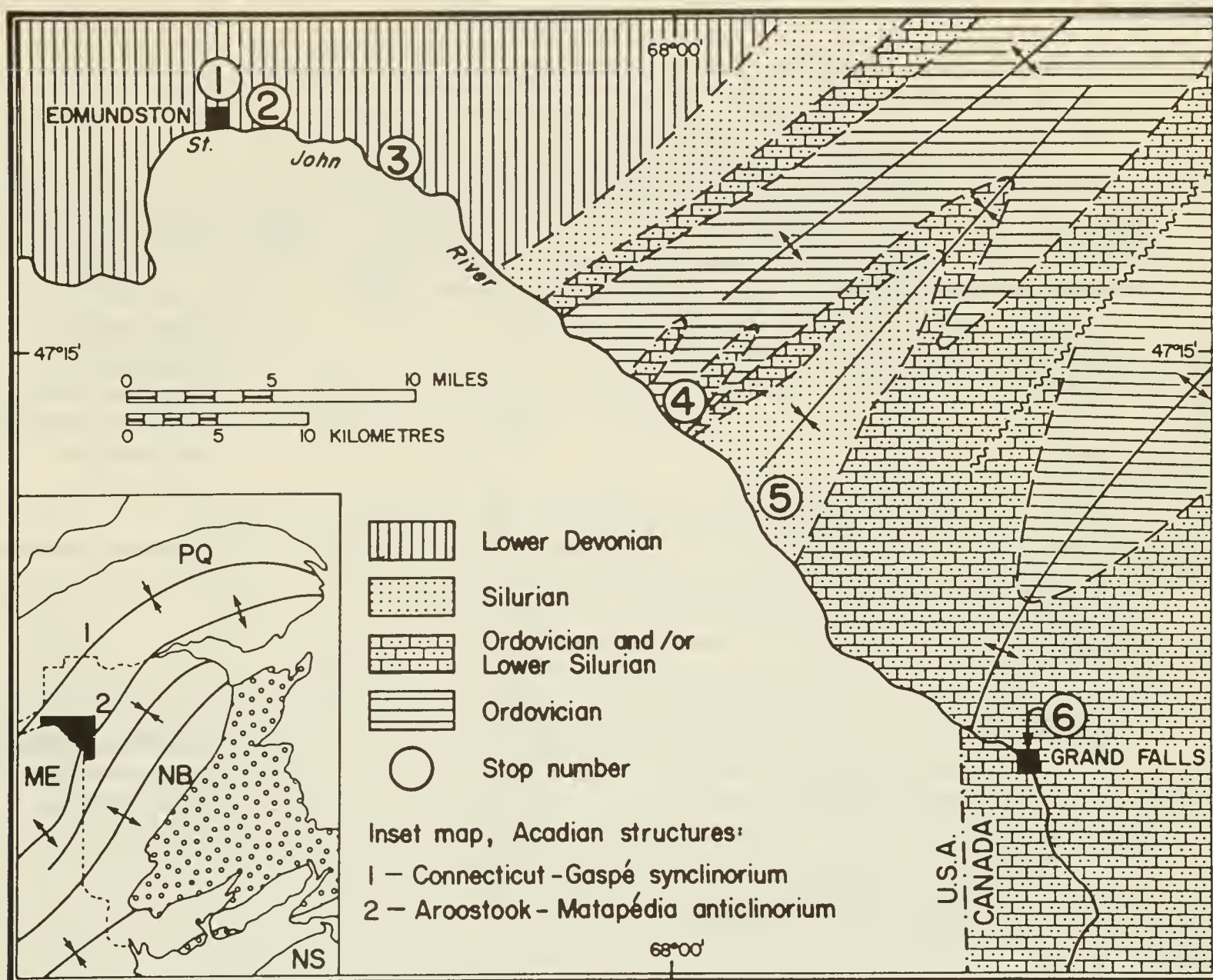


Figure 1. Generalized geological map of the Edmundston - Grand falls area, New Brunswick (based on St. Peter, 1977, fig. 2). Inset map shows Acadian (Devonian) anticlinoria and synclinoria in New Brunswick and adjacent Maine and Quebec (based on Rast and Stringer, 1974), the Carboniferous cover (ornament), and the location of the trip area.

dipping subduction zone within a former single continental plate. Whatever the cause, the folding and cleavage attest to very significant shortening during the Acadian (northern Appalachians) and Erian (Ireland to Scandinavia) orogeny. In the Edmundston - Grand Falls area, Middle to Upper Ordovician as well as Silurian-Lower Devonian sediments were involved in the Acadian orogeny. Taconian deformation is lacking in the Carys Mills Formation; some folding in the underlying unnamed unit may be Taconian (Hamilton-Smith, 1969, 1970).

	SERIES	Siegas Hamilton - Smith, 1970		AROOST. - MAT. ANTICLINORIUM St. Peter, 1977	
				EAST	WEST
LOWER DEVONIAN	Emsian				<u>Temiscouata Formation</u> siltstone, slate, arkose, greywacke, conglomerate
	Sieg.				
	Ged.				
SILURIAN	Ludlow	? — ? —			<u>Perham Formation</u> slate, sandstone, lithic wacke, limestone
		Perham Formation	<u>Upper Member</u> shale, siltstone, minor sandstone		
	<u>Lower Member</u> calcareous slate, minor siltstone				
	Wenlock				
			Landover		
		<u>Siegas Formation</u> sandstone, slate, limestone		<u>Upsalquitch Fm.</u> siltstone, sandstone, greywacke	
ORDOVICIAN		<u>Carys Mills Fm.</u> limestone, slate	<u>Matapedia Group</u> limestone, shale		
		<u>Unnamed Unit</u> slate, shale, sandstone	<u>Grog Brook Group</u> argillite, sandstone, greywacke; minor conglomerate, limestone		

Table 1 Lithology and stratigraphic correlation of Ordovician to Lower Devonian formations in the Edmundston-Grand Falls area, New Brunswick.

Folding and cleavage

The tight upright Acadian folds in the Middle Ordovician-Lower Devonian sediments of the Edmundston - Grand Falls area trend NE-SW and are similar in style to Acadian folds throughout much of New Brunswick and adjacent regions of the northern Appalachians. Sporadic to well developed cleavage associated with the folds is persistently steep or vertical. The folds and cleavage have been designated as F_1 and S_1 , and the cleavage has been described as axial planar to the folds (St. Peter, 1977). Stereographic plots of bedding and cleavage (Hamilton-Smith, 1970, figs. 10-14) suggest, however, that the cleavage strikes a few degrees clockwise of the strike of the fold axial surfaces. Cleavage noncoplanar with fold axial surfaces is a widespread feature of Acadian deformation in the northern Appalachians (Stringer, 1975).

The S_1 cleavage, recorded as a slaty or fracture cleavage by St. Peter (1977), is formed by parallel or anastomosing partings, spaced at about 10 mm intervals in sandstone beds and much more closely in siltstone and mudstone (shale) beds. In thin sections of siltstones and mudstones at the hinges of F_1 folds, the mesoscopic S_1 cleavage partings are seen as dark films (c. 0.005 mm thick) made of fine-grained white mica, opaque minerals and irresolvable material, in which the micas are aligned subparallel or parallel to the cleavage direction. In the microlithons (c. 0.04 mm thick) between the cleavage films, fine-grained white mica, quartz and calcite grains commonly show no obvious preferred orientation, apart from occasional parallelism with bedding. Diagenetic chlorite grains within the microlithons often display mineral cleavage planes oriented perpendicular or at a high angle to the S_1 cleavage films, and muscovite laths are occasionally intergrown with the chlorite grains parallel to the mineral cleavage of the chlorite host. Compression within the microlithons perpendicular to the S_1 cleavage is indicated by crenulation of the chlorite mineral cleavages and the intergrown muscovite laths. Where thin calcareous or quartzose siltstone beds are present within the folded mudstones, the S_1 cleavage films converge towards re-entrants in the silty layers formed by buckles and by reverse microfaults (Figs. 2 and 4). The S_1 cleavage films correspond to pressure solution planes (Durney, 1972; Williams, 1972; Gray, 1979; Stephens *et al.*, 1979), and the cleavage should therefore be described as a pressure solution cleavage. The pressure solution cleavage films represent a significant component of the shortening normal to the cleavage direction (Gray, 1979, pp. 114-7).

Evidence for the origin of the S_1 cleavage films by pressure solution is provided by a progressive development of microstructures (Gray, 1979, fig. 3) which can be recognized in thin sections of the S_1 cleavage from the hinges of F_1 folds in laminated silty mudstones in the Edmundston - Grand Falls area (Fig. 3). Shortening in the mudstone is represented by uniformly and closely spaced pressure solution S_1 cleavage films, and by compression of the intervening microlithons indicated by the crenulated chlorite grains. Shortening in the interbedded thin silty layers, in which pressure solution films commonly are lacking, occurred by symmetric to asymmetric buckling concomitant with development of the pressure solution films in the mudstone. Adjacent asymmetric buckle folds may

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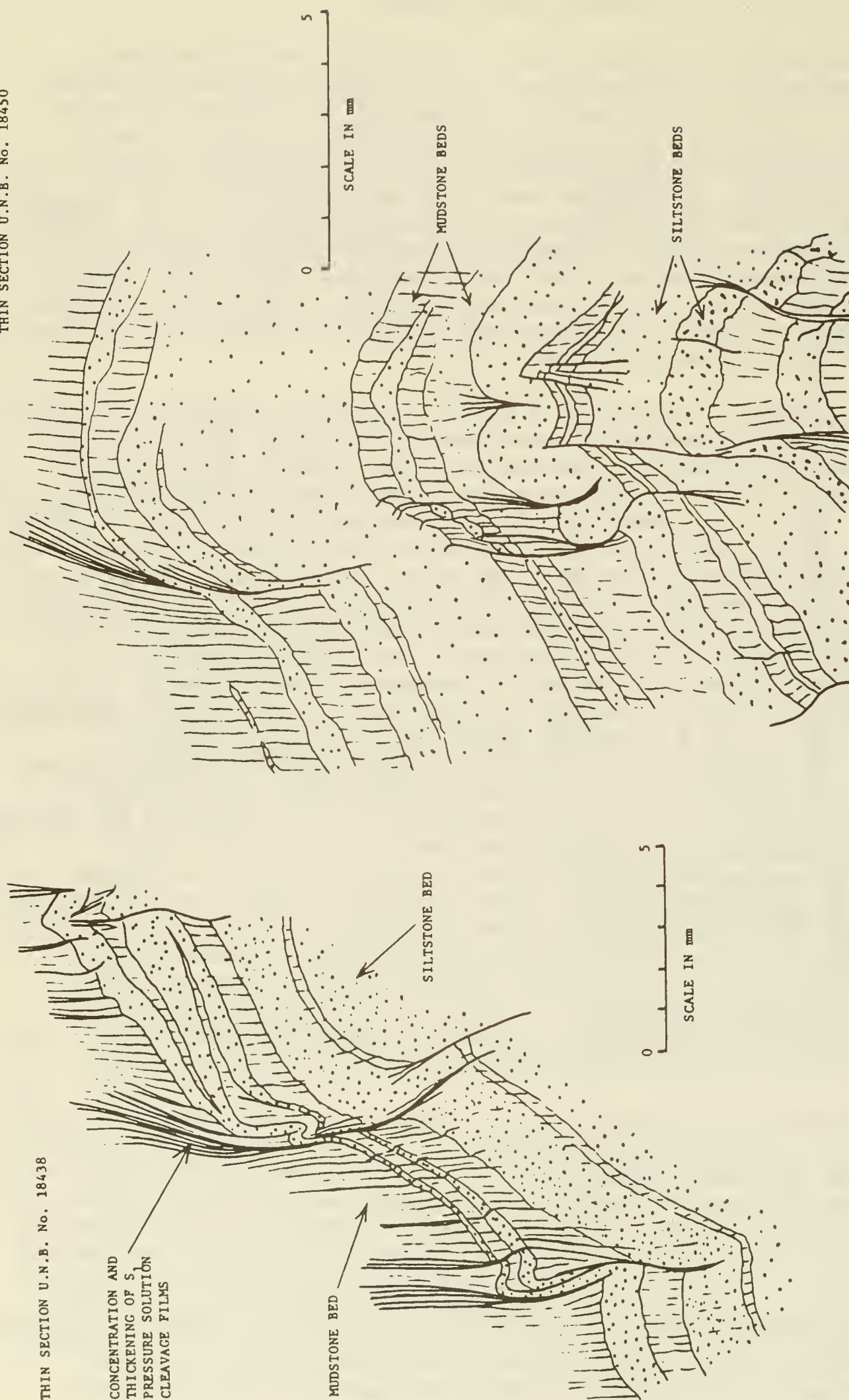


Figure 2. Concentrated and thickened S_1 pressure solution cleavage films in mudstone beds converge upon the short limbs of asymmetric microfolds in thin siltstone beds. Silurian Perham Formation, St. Leonard, New Brunswick (Stop 5).

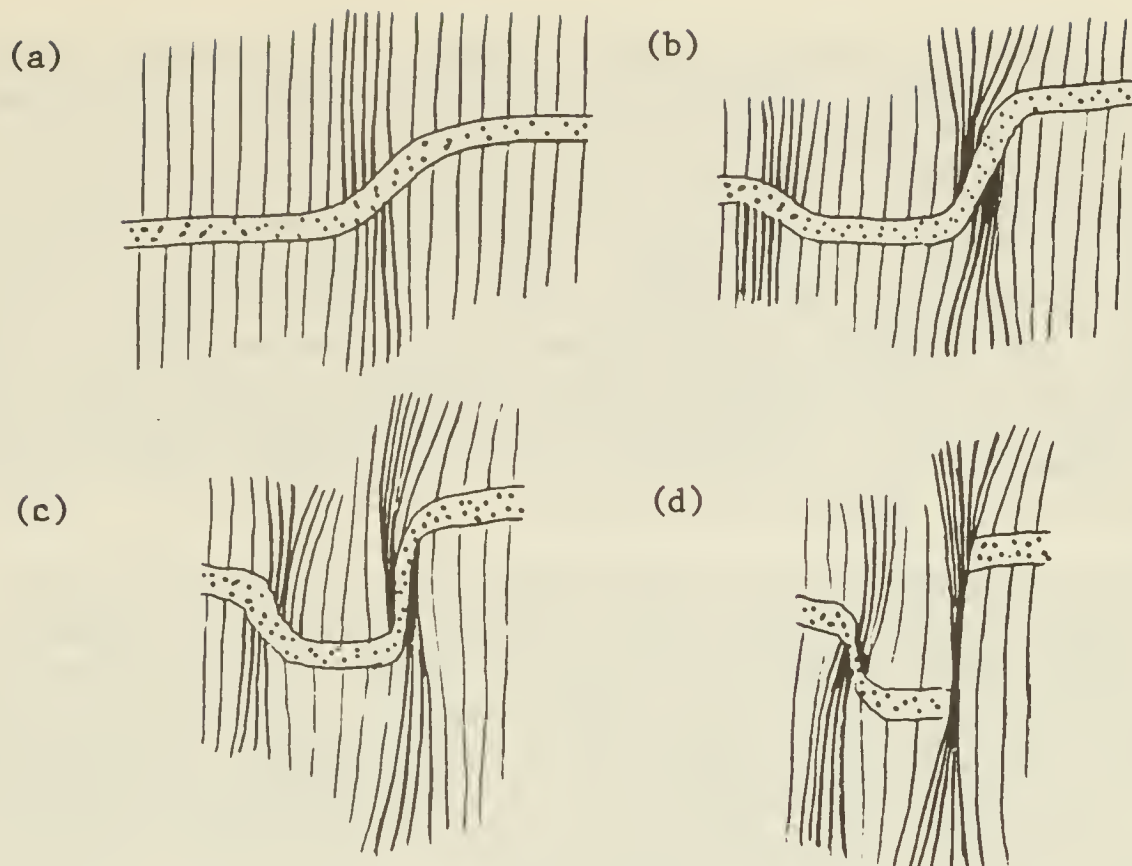


Figure 3. Schematic diagram illustrating the progressive development of microstructures in laminated silty mudstone during pressure solution on S_1 cleavage planes in the mudstone concomitant with asymmetric buckling of a thin silty layer. (a) Evenly spaced S_1 pressure solution films in the mudstone show slight concentration and thickening adjacent to short limb of gentle asymmetric microfold in the silty layer. (b) and (c) Progressive concentration, thickening and convergence of S_1 pressure solution films against the rotating and partly dissolved silty layer in the short limb, and development of a second asymmetric microfold with opposite vergence. (d) Complete solution of short limb produces apparent offset of silty layer along thickly concentrated S_1 pressure solution films.

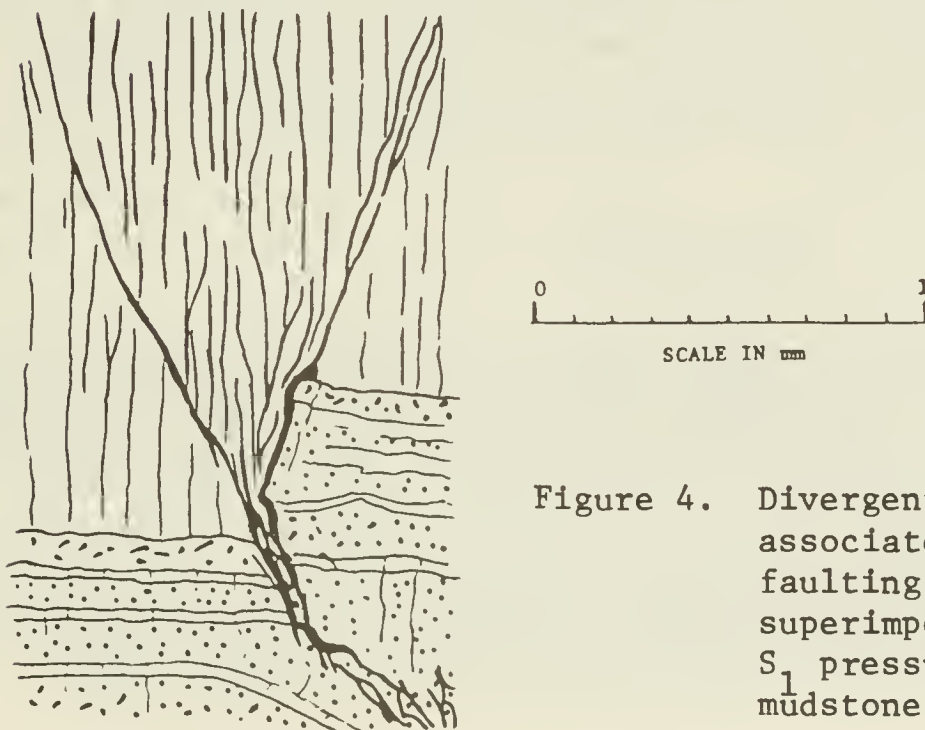


Figure 4. Divergent pressure solution films associated with reverse micro-faulting of a siltstone bed are superimposed upon uniformly spaced S_1 pressure solution films in the mudstone bed.

show opposing vergence. Pressure solution films in the mudstone are more closely spaced and slightly thicker near the short limbs of the asymmetric folds, indicating more intense solution of the microlithons (Fig. 3a). With progressive rotation of the short limbs, the pressure solution films in the mudstone become more concentrated, converging towards each side of the short limb (Fig. 3b), and the silty layer is progressively dissolved (Fig. 3c). The final stage is marked by loss of the short limb and the development of a continuous zone of concentrated pressure solution films along which the silty layer appears offset, resembling a microfault (Fig. 3d). The different stages of microstructural development can often be traced across several silty layers within a single thin section, from gentle asymmetric microfold to apparent microfault.

Shortening of the thin silty layers by actual microfaulting is indicated in places by low to high angle reverse microfaults, commonly associated with the short limbs of asymmetric microfolds. Concentrations of pressure solution films which diverge away from the reverse fault in the silty layer into the adjacent mudstone may be superimposed upon uniformly spaced pressure solution films formed in the mudstone prior to the microfaulting (Fig. 4). Thus polyphase pressure solution microstructures may be produced during a single phase of shortening.

Slump folding

Recumbent folds which pre-date the F_1 folding and the S_1 cleavage are common in the Ordovician-Lower Devonian sediments of the Edmundston - Grand Falls area (Stops 2 and 6). Acadian folding of the earlier folds is prominent in the Lower Devonian Temiscouata Formation (Stop 3), and has been described in the Lower Silurian Siegas Formation (Hamilton-Smith, 1970, pp. 29-30). The earlier folds exhibit no constancy in scale, style or orientation, and lack axial plane cleavage (St. Peter, 1977), and have been interpreted as slump folds in unconsolidated water-laden sediments. Large scale recumbent pre- F_1 folds indicated by inverted bedding in the Upper Ordovician-Lower Silurian Carys Mills Formation in the Woodstock area (Rast *et al.*, 1980) 110 km south of Grand Falls have been interpreted by Rast as soft-sediment deformation features due to submarine slip, although Rast noted (*ibid.*, p. 7) that it is very strange that no soft sediment brecciation is associated with the large scale structures.

Sedimentology

Apart from the absence of late Silurian to early Devonian (Pridolian to Gedinnian) strata, which St. Peter (1977) interprets as a Ludlovian to Gedinnian disconformity, sedimentation within the area between Edmundston and Grand Falls was essentially continuous from at least Middle Ordovician (zone of *Nemagraptus gracilis*) to Lower Devonian (Emsian) time. This sedimentation occurred within a narrow northeast/southwest elongate trough extending from Gaspé through New Brunswick and into Maine and termed the Central Clastic Belt and/or Aroostook-Matapedia Carbonate Belt by Ayrton *et al.* (1969). Strata in this belt have been assigned various group and/or formational names at different places along the belt.

Within the Edmundston and Grand Falls section of the belt, relatively few and detailed sedimentological analyses have thus far been undertaken, notable exceptions being the studies of Hamilton-Smith (1970, 1971a, 1971b) in the Siegas area and St. Peter (1977) in the Edmundston-Grand River area. It is apparent from these studies that the Middle Ordovician-Lower Devonian strata were deposited in a deep-water trough, within which a series of isolated uplifted areas provided sedimentary detritus. Evidence of deep-water deposition is not only provided by detailed analysis of sedimentary lithofacies and depositional mechanisms but also by the existence of a deep-water *Nereites* ichnofacies recently described by Pickerill (1980) immediately to the northeast in the Matapedia district in northern New Brunswick.

Sedimentological aspects emphasized on this excursion will be directed essentially toward the Late Ordovician-early Silurian Carys Mills (Stop 6) and the early Llandovery Siegas (Stop 4) Formations (Table 1). The Siegas Formation is in part clearly erosive into the underlying Carys Mills Formation, the estimated depth of erosion being in the order of c. 25-50 m (Hamilton-Smith, 1971b). Proximal turbidites with partial and/or more complete Bouma sequences and well-developed resedimented limestone conglomerates characterize these erosive fills. Associated massive and structureless sandstone beds most probably represent deposition from grain and/or fluidized flows. Proximal-distal relationships and palaeocurrent analysis undertaken by Hamilton-Smith (*ibid.*) suggest derivation from the north and north-west from an isolated topographic high. The Siegas Formation was thus a localized channel sequence developed within the Carys Mills Formation. More distal lithofacies of the former are undoubtedly lateral facies equivalents of the latter.

The Carys Mills Formation and lateral equivalents throughout the belt (*e.g.* Smyrna Mills Formation, Matapedia Group, *etc.*) consist essentially of thinly interbedded calcareous argillites and argillaceous calcitic and ankeritic limestones with thin and randomly developed graded calcarenites with a substantial quartz component. Differential solution has commonly etched the more limy beds to a greater extent so that the rocks have a markedly ribbed appearance and are commonly referred to as 'ribbon limestones' (Ayrton *et al.*, 1969). Previous authors (*e.g.* Ayrton *et al.*, 1969; St. Peter, 1977) have interpreted the Carys Mills Formation as having resulted from deposition by turbidity currents in a distal palaeoenvironment. We believe, however, that this is an oversimplification and that whilst some units are suggestive of deposition from turbidity currents, the majority of units are not. The thinly bedded calcarenites exhibit partial Bouma sequences, particularly T_a and T_{a-b} sequences, are often erosive and contain broken and abraded shallow-water benthic marine fossils. We believe these units to be the result of deposition by turbidity currents. The thinly bedded calcareous argillites and argillaceous calcitic and ankeritic limestones are, however, considered to result from deposition by hemipelagic and normal bottom-following contour currents *viz.*:- contourites. Though the differentiation of contourites and turbidites in the geologic record is extremely difficult and often hazardous because of the lack of suitable distinguishing criteria (Bouma and Hollister, 1973), the following observations are suggestive of

deposition by deep thermohaline boundary (contour) currents which paralleled the ancient palaeoslope.

(i) The presence of abundant intraformational slump horizons within the Carys Mills Formation together with minor resedimented debris flow and olistostromal horizons. Such an association is more consistent with slope (both modern and ancient) environments rather than distal or overbank turbidite associations.

(ii) The widespread occurrence of similar facies which parallel the presently exposed margins of the trough.

(iii) The presence of distinctive submarine channel lithofacies which cut the regional strike and palaeocurrent (*e.g.* the Siegas Formation) of the Carys Mills Formation.

(iv) The presence of a trace-fossil assemblage more consistent with slope deposition rather than deep bathyal or abyssal palaeoenvironmental conditions (*e.g.* amongst others, vertical burrows which cross-cut successive units, *Diplichnites*, *Fucusopsis*, *Gyrochorte* and spreiten-bearing forms).

(v) The presence of lenticular and irregular, commonly bioturbated, horizons comparable to the 'muddy contourites' of Stow and Lovell (1979).

It is therefore suggested that the Carys Mills Formation represents a succession of contourites and hemipelagic deposits which include randomly and thinly developed turbidites. In modern regimes this association is most consistently developed in continental rise environments (Stow and Lovell, 1979). Nevertheless, as outlined above, evidence in the Carys Mills Formation suggests that such an association may be equally as common in an ancient slope environment.

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Itinerary

Assembly point is in the parking lot of Keddy's Motor Inn, Presque Isle. Starting time 8:00 AM (Eastern Standard Time). Upon leaving the parking lot, drive out of Presque Isle and follow Route 1 for 60 miles (96 km) north via Van Buren and Madawaska to Edmundston. Proceed through Canada Customs and Immigration at the international border between Madawaska, Maine and Edmundston, New Brunswick, and drive 2 miles (3.2 km) north through Edmundston East to the Edmundston Motel on Route 2 (Trans-Canada Highway).

Mileage Km

00.0 00.0 Stop 1. Park on southwest side of Trans-Canada Highway (Route 2) at front of Edmundston Motel, Edmundston. Cross highway to 100 m long road section.

Lower Devonian Temiscouata Formation. Dark grey laminated shaly siltstones are deformed by symmetrical NE-SW trending Acadian open folds (half-wavelength 50 m) and subvertical S_1 pressure solution cleavage striking 052° . Cleavage/bedding¹ intersections plunge NE between 10° and 40° . At the northwest end of the road section, small scale (0.5 cm) load casts in thinly bedded fine grained sandstone interbeds in the hinge of a syncline indicate that beds are the right way up. About half way along the road section, spaced composition banding parallel to vertical cleavage in gently inclined beds indicates differential mineral concentration along cleavage planes by pressure solution.

Return to vehicles and continue east on Route 2.

2.3 3.7 Stop 2. Park on south side of highway at roadside outcrop 30 m west of overhead power transmission lines.

Slump structures in Temiscouata Formation. At the west end of the outcrop, tight minor recumbent slump folds are exposed at the top of the rock faces within gently inclined bedding. At the east end of the outcrop, a 1 m thick horizon of irregular slump structures dips SE within the limb of an upright NE-SW trending Acadian fold with subvertical S_1 pressure solution cleavage which strikes 053° and cuts through the slump structures. In places, a weak second cleavage, S_2 , is present at a slight angle to the S_1 cleavage.

Return to vehicles and continue east on Route 2.

5.0 8.0 St. Basile

7.3 11.7 Stop 3. Park on south side of highway by the Weigh Station sign. Cross highway to outcrop on north side of road.

Mileage Km

Folded larger scale recumbent slump fold, Temiscouata Formation. A tight recumbent slump fold in dark grey laminated shaly siltstone exposed for 20 m along its axial surface is folded around the hinge of a NE-SW trending Acadian anticline. S_1 pressure solution cleavage inclined steeply NW cuts through both limbs of the slump fold. Sedimentary structures in the core of the anticline indicate that bedding in the lower limb of the slump fold is inverted. There is no fabric parallel to the axial surface of the slump fold. Minor isoclinal slump folds verging southeast are discernible on the upper limb of the larger scale slump fold near the east end of the outcrop.

Return to vehicles and proceed east on Route 2.

- 11.3 18.2 Cross bridge over Green River (Riviere Verte).
- 17.3 27.8 Ste. Anne de Madawaska
- 19.9 32.0 Turn left off Route 2, follow gravel road round back of house and proceed east on old paved road.
- 20.2 32.5 Stop 4. Park on road near gates to Siegas Quarry. Walk 300 m north along road into quarry.

Lower Silurian Siegas Formation. The Siegas Quarry is located on the northwest flank of the Aroostook-Matapedia Carbonate Belt of Ayrton *et al.* (1969), the most prominent large structure of northeastern Maine and northwestern New Brunswick during the Taconic orogeny (Hamilton-Smith, 1971a). Although borderlands on both sides of this belt were deformed, uplifted and eroded during the Taconic orogeny, the Aroostook-Matapedia Carbonate Belt continuously received sediment so that no unconformity of any regional significance was developed within this belt during Middle Ordovician-Late Silurian times. The characteristic rocks of this part of the belt are known regionally as the Carys Mills Formation, a succession of calcareous thinly bedded flysch containing graptolites of Middle Ordovician (*Orthograptus truncatus* var. *intermedius*) to early middle Llandovery (*Monograptus communis*) age. The Siegas Quarry, however, exposes rocks of a locally developed clastic facies within this more regionally developed calcareous flysch known as the Siegas Formation. This formation ranges in thickness from c. 105-240 m in the Siegas district and thins to zero within 10 km to the south (Hamilton-Smith, 1971a).

The Siegas Quarry represents the principal reference section of the Siegas Formation (Hamilton-Smith, 1970) and exhibits

Mileage Km

excellent exposures of limestone conglomerates, sandstones, siltstones and shales. Fragmentary brachiopod genera include *Stricklandia*, *Plectothyrella*, *Leangella*, *Mendacella*, *Eoplectodonta*, *Dalmanella* and *Protatrypa* and attest to an early Llandovery age. These and additional faunas are listed in Ayrton *et al.* (1969).

The limestone-conglomerates and sandstones exhibit features indicative of an origin by turbidity current grain support mechanisms. These beds, which range in thickness from 4 cm - 8 m, show excellently developed clast imbrication, partial and more complete Bouma sequences and a whole variety of sole structures, such as gutter casts, prod, bounce and other tool markings, flute casts, load casts, ripple marks and dewatering features. Detailed palaeocurrent analysis by Hamilton-Smith (1971a) indicated a provenance from the north and northwest, which he suggested to be an isolated relatively discrete uplift similar to others previously described in northeastern Maine by Pavlides *et al.* (1964). The interbedded siltstones and shales are generally 1-100 cm thick and exhibit parallel or more rarely cross-laminations. They represent material reworked and/or deposited by normal bottom marine currents. Organic activity is evidenced by the obscure trails preserved on the upper surfaces of some siltstones and shale beds as well as the more recognizable ichnogenera *Chondrites*, *Planolites*, *Fucusopsis* and *Diplichnites*. Comparison of the Siegas Facies association with similarly described ancient analogues suggests deposition in a submarine channel or slope/base of slope environment. The localized development of the formation suggests that a channel is probably more realistic.

More detailed and complete descriptions of the Siegas Formation lithofacies and regional relationships may be obtained in Hamilton-Smith (1969, 1970, 1971a, 1971b).

Return to vehicles and continue east along old paved road.

20.5 33.0 Turn left onto Route 2 and continue eastward.

25.0 40.3 Stop 5. Park on south side of highway at roadside outcrop 600 m west of Route 2/Route 17 interchange, St. Leonard.

Silurian Perham Formation. A NE-SW trending Acadian syncline plunging gently SW at the east end of the outcrop deforms slightly calcareous grey laminated silty mudstones. Intense small-scale buckling of thin (1-5 mm) siltstone beds indicates considerable shortening concomitant with pressure solution on the S_1 cleavage planes in the mudstone beds (Fig. 2). The S_1 cleavage dips steeply SE. Convergent cleavage refraction₁ is prominent in the thicker (3 cm) silty beds.

Mileage Km

Return to vehicles and continue east on Route 2.

- 32.6 52.5 Cross bridge over C.N.R. tracks.
- 36.1 58.1 Turn right off Route 2 at interchange exit for Grand Falls.
Cross bridge over Route 2 and follow road down to Grand Falls.
- 38.0 61.2 Stop 6. Turn right off road into parking area of Falls View Motel, Grand Falls. Clamber down bank to extensive outcrop at confluence of Saint John River and Little River.

Upper Ordovician-Lower Silurian Carys Mills Formation, Aroostook-Matapedia anticlinorium. Beds striking NE-SW and dipping steeply SE young towards the southeast and are intersected by S_1 pressure solution cleavage which strikes ENE-WSW and dips $45-85^\circ$ NNW. The c. 100 m thickness of beds exposed includes numerous horizons of spectacular intraformational slump folding. In places, the axes of tight slump folds and convolute structures plunge steeply SE at a maximum down the dip of the beds, but most of the slump folds are variably oriented.

The majority of the beds exposed here are composed of the more typical 'ribbon limestones' of the Carys Mills Formation. The argillaceous calcitic and ankeritic limestones are internally quite variable, some exhibiting grading, some parallel lamination, some cross-lamination, whilst others exhibit bioturbation or appear to be generally structureless. The majority of beds are laterally continuous whilst others are lenticular and irregular. Associated thin and graded calcarenites may also be observed, many of which have been completely dismembered due to *in situ* soft-sediment deformation. Without doubt, however, the most obvious soft-sediment deformation features are the spectacularly developed intraformational slump fold horizons, some of which are 2 m in thickness. At the southeastern edge of the exposure occurs an enigmatic debris flow or olistostromal horizon. The presence of this and the slump folds attest to the presence of a slope during deposition of the ribbon limestones.

Return to vehicles and drive out of motel car park, turning right onto the road.

- 38.2 61.5 Turn right (south) following sign for Grand Falls Centre and Fredericton, crossing bridge over the Saint John River. Continue south along Broadway through Grand Falls centre.
- 38.8 62.4 At south end of Broadway turn left, following sign for Route 2 East. Follow road up the hill out of Grand Falls.

Mileage Km

- 41.1 66.2 Turn left onto Route 2 at Grand Falls Portage, following sign for Fredericton, Route 2 East. Proceed south on Route 2.
- 54.0 86.4 Stop 7 (optional). East side of highway (Route 2), north of Four Falls.

Pre-Acadian (Silurian?) dykes in 'ribbon limestones' of the Upper Ordovician-Lower Silurian Carys Mills Formation.

Return to vehicles and continue south on Route 2.

- 59.8 95.7 Turn right off Route 2 at Perth-Andover interchange onto Route 19 (N.B.)/Route 167 (Maine) and drive west to Fort Fairfield, passing through U.S. Immigration and Customs at the international border. Follow Route 167 from Fort Fairfield to Presque Isle. The distance back to Presque Isle from Perth-Andover is 18 miles (29 km). End of field trip.

STRATIGRAPHIC AND STRUCTURAL RELATIONS IN THE TURBIDITE SEQUENCE OF SOUTH-CENTRAL MAINE

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Introduction

The turbidite section of south-central Maine occupies the southeast part of the Merrimack synclinorium which extends from the vicinity of Houlton, Maine, through southern Maine, central New Hampshire, central Massachusetts, and into central Connecticut. These rocks belong to a thick (>5000 m) section of Silurian age that is in contact to the east with a terrane of plagioclase gneiss and amphibolite of Precambrian (?) or early Paleozoic age. Early foliated plutons, presumably of pre-Silurian age, intrude the gneiss-amphibolite terrane. Devonian plutons of granite, quartz monzonite, and granodiorite cut both the turbidite and gneiss-amphibolite terranes. The gneiss amphibolite terrane was metamorphosed to epidote-amphibolite facies in Precambrian (?) or early Paleozoic time. Both the turbidite and gneiss-amphibolite terranes were metamorphosed in a Buchan-type facies series in Early Devonian time. The Buchan-type metamorphism increases in grade to the south.

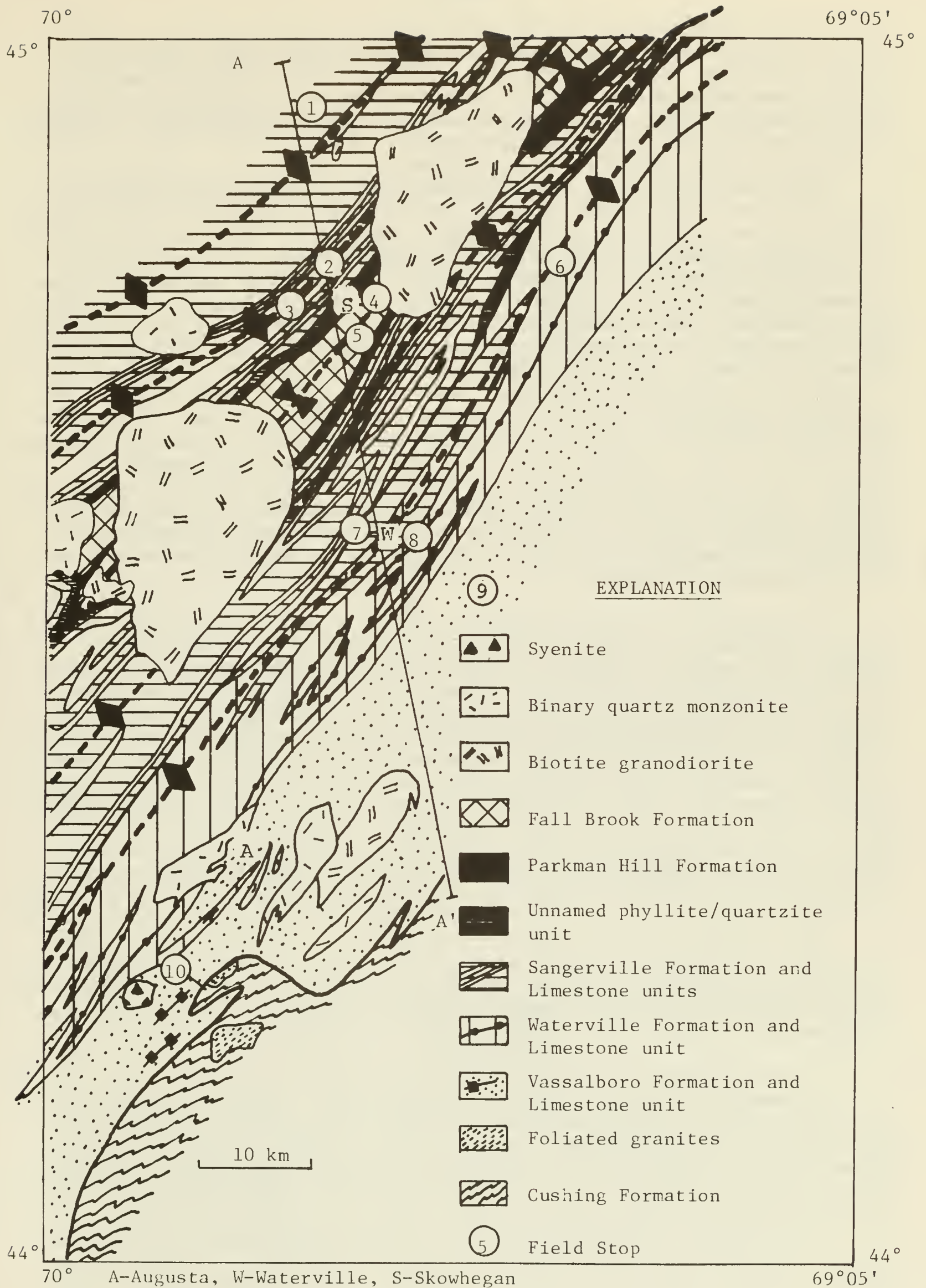
The distribution of rock units is shown in Figure 1. The region includes the Skowhegan, Pittsfield, Norridgewock, Waterville, Vassalboro, Augusta, and Gardiner 15' quadrangles of the U.S. Geological Survey Atlas of Topographic Maps. The geology is based on work of Barker (1961), Osberg (1968), Heinonen (1971), Griffin (1973), Ludman (1976, 1977), Pankiowskyj, and others (1976), and on unpublished notes of Pankiowskyj, Newburg, and Osberg.

Stratigraphy

The stratigraphic column is presented in the explanation of Figure 1. The paleontological control comes from Osberg (1968) and from the data of Pankiowskyj and others (1976).

The relative stratigraphic position of lithic units has been determined by primary sedimentary features. But because of complex structural relations and because both upward-facing and downward-facing sections have been recognized, primary features are most relevant at or near formational contacts. This circumstance limits the number of critical observations.

Cushing Formation. The Cushing Formation is exposed along the east boundary of the area shown in Figure 1. These rocks are continuous southward to the latitude of Portland, where the name Cushing was first used (Katz, 1917; Hussey, 1971). The Cushing Formation in south-central Maine consists of three divisions: (1) a lower quartz-K feldspar-plagioclase-biotite gneiss consisting of quartzofeldspathic layers 1-3 cm thick alternating with thinner biotite-rich layers, (2) a medial plagioclase-quartz-biotite gneiss



A-Augusta, W-Waterville, S-Skowhegan

Figure 1. Geologic map of south-central Maine.

interlayered with biotite-bearing amphibolites, and (3) an upper, somewhat massive plagioclase-quartz-biotite granulite. A rusty zone, forming a nearly continuous unit along the west boundary of the Cushing Formation, may be a mineralized zone along a thrust contact or it may partly represent a stratigraphic unit. All sequences contain abundant pegmatites. No thickness can be given for the Cushing Formation.

Age relations for the Cushing Formation are not well established. Mineralogy and texture from the north extremity of its outcrop indicate that it has been metamorphosed to higher grade than the overlying Silurian turbidite section. The Cushing Formation is probably early Paleozoic in age or possibly Precambrian (?). The lithology and stratigraphic position of the Cushing Formation are similar to those of parts of the "Massabessic" gneiss in New Hampshire where Besancon and others (1977) have determined a 600 m.y. age and to those of the Nashoba Formation in Massachusetts. The Cushing Formation, however, cannot be traced continuously into these units.

Vassalboro Formation. The name Vassalboro sandstone was coined by Perkins and Smith (1925) for massively bedded, bluish graywacke cropping out between the Kennebec River and China Lake. The name has since been amended to Vassalboro Formation (Fisher, 1941; Barker, 1961) because of the variability of the unit. The name Kenduskeag Formation as used by Ludman and Griffin (1974) has been applied to some rocks originally included within the Vassalboro Formation and refers to a distinctive lithology, but it may not represent a mappable unit.

The Vassalboro Formation Consists of bluish gray, slightly calcareous, quartz wacke and quartz-mica phyllite/schist. The proportion of these lithologies varies from place to place, but quartz wacke always makes up a part of any outcrop. Thickness of bedding varies widely; the quartz wacke is present in beds that range from 7 cm to several meters in thickness, whereas the schist/phyllite is present in beds that range from a few millimeters to several centimeters in thickness. Beds of quartz wacke have parallel lamination and, locally, cross lamination and convolute tops. At higher metamorphic grades, the quartz wacke contains diopside, green amphibole, clinozoisite, and abundant plagioclase. A prominent unit of quartz-mica schist and gray marble observed at several localities near the east boundary of the Vassalboro Formation may represent inliers of Waterville Formation or a heretofore unrecognized stratigraphic unit within the Vassalboro.

The Vassalboro Formation has a large breadth of outcrop even though attitudes of bedding are mostly vertical throughout the area. Consequently, estimating the thickness of the Vassalboro is difficult. Its minimum breadth of outcrop and its maximum thickness is approximately 3,200 m, but presumably it is less than this figure due to folding within the unit.

A single locality containing identifiable fossils has been described in the Vassalboro Formation (Pankiwskyj and others, 1976). Graptolites from this locality suggest a Llandoveryan to Ludlovian Age. Considering the thickness of the Vassalboro Formation and the ages of the overlying formations, I think that the Vassalboro is probably Llandoveryan, although the lower part of the formation could be as old as latest Ordovician.

The Vassalboro Formation is equated to the lower part of the Smyrna Mills Formation of northeastern Maine, possibly part of the Berwick Formation in southern Maine, the Oakdale Quartzite and the Paxton Quartz Schist in central Massachusetts, and the Hebron Formation in Connecticut. It may

correlate with the Quimby Formation in western Maine.

Waterville Formation. Perkins and Smith (1925) used the name Waterville shale to designate rocks exposed west of the Kennebec River in the vicinity of Waterville. Osberg (1968) changed the name to Waterville Formation, which included an eastern and a western facies. Present usage restricts the Waterville Formation to the eastern facies; the western facies is included in the Sangerville Formation (Ludman, 1976).

The Waterville Formation is dominantly a thinly laminated phyllite/schist. Typically, beds of quartzite or slightly calcareous quartzite alternate with beds of quartz-mica phyllite/schist. Bedding thicknesses are generally less than 3 cm. Convolute structures in the quartzitic layers and grading between quartzitic and phyllitic layers can be observed locally. The phyllite/schist layers contain a variety of mineral assemblages depending on metamorphic grade. Assemblages containing combinations of quartz, plagioclase, muscovite, biotite, chlorite, garnet, staurolite, cordierite, andalusite, and sillimanite are common.

A conspicuous unit composed of gray limestone and quartz-mica phyllite forms an excellent mapping horizon within the Waterville Formation. This unit is thinly bedded with limestone layers 1-6 cm thick alternating with phyllite layers generally less than 4 cm thick. At high metamorphic grades, this unit contains diopside, grossularite, green amphibole, phlogopite, and plagioclase.

The thickness of the Waterville Formation cannot be determined from a single section because of the complexity of folding. However, an estimate of maximum thickness can be determined by measuring the narrowest breadth of outcrop from its base to the limestone member and adding that value to the minimum breadth of outcrop between the limestone member and the upper boundary of the formation. The resulting figure is 1,100 m. Of course, the thickness of the formation may vary geographically along and across strike.

Three fossil localities in the Waterville Formation have been described (Osberg, 1968; Pankiowskyj and others, 1976). Two of these localities contain fossils that range in age between Ordovician and Early Devonian, but the third contains fossils that limit the age to a range of late Llandoveryan to Ludlovian.

The Waterville Formation may be correlative to the west with the sequence Turner through Anasagunticook Formations as used by Pankiowskyj and others (1976) and the Greenvale Cove Formation. It may correlate to the south with at least part of the Eliot Formation.

Sangerville Formation. The Sangerville Formation includes exposures of graywacke and phyllite in the Skowhegan quadrangle (Ludman and Griffin, 1974; Pankiowskyj and others, 1976). These rocks are in part continuous with rocks that Osberg (1968) interpreted as the western facies of the Waterville Formation, and the name, Sangerville Formation, is adopted for these rocks. This unit also contains rocks that Osberg (1968) called the Mayflower Hill Formation, a name which should now be abandoned.

The Sangerville Formation lies stratigraphically above the Waterville Formation, and as used here consists of two principal lithologies: a lower graywacke-phyllite unit and an upper feldspathic wacke and interbedded limestone unit. The upper feldspathic wacke and interbedded limestone unit makes up the bulk of the formation.

Slightly calcareous graywacke and quartz-mica phyllite are the common rocks of the lower unit. Beds range from 7 cm to a meter in thickness, and the thickness of the graywacke is eight or nine times that of the phyllite. Beds are commonly graded. Parallel lamination is common and delicate cross-beds can be seen locally. A black sulfidic quartz-mica phyllite is found discontinuously at the contact with the Waterville Formation.

The upper unit consists principally of light-gray feldspathic graywacke and medium-gray quartz-mica phyllite/schist. Some of the graywacke is slightly calcareous. Beds are 8 cm to 1.5 m thick, and, although most beds are ungraded, they locally show grading, cross lamination, and scour-and-fill features. The graywacke is locally conglomeratic and contains clasts of slate, feldspar, plutonic rocks and volcanic rocks. The phyllite/schist is laminated to massive. At appropriate metamorphic grades, it contains staurolite, garnet, and sillimanite.

Two conspicuous units of gray limestone are present in the upper unit. Lithologically, the limestone is similar to the limestone member of the Waterville Formation, a circumstance that has caused much confusion in earlier studies (Osberg, 1968; Ludman and Griffin, 1974; and Pankiowskyj and others, 1976).

The thickness of the Sangerville Formation is difficult to estimate, and the writer has found no good place to reconstruct its thickness. Ludman (1977) gave a thickness in excess of 2,000 m.

Fossils are found at three localities in the area shown in Figure 1. Graptolites from these localities indicate an age from late Llandoveryan to Wenlockian (Osberg, 1968; Pankiowskyj and others, 1976). This age range is confirmed by fossil information from adjacent areas.

The Sangerville Formation is equivalent to the Rangeley Formation in western Maine. No correlatives are known in southern Maine.

Unnamed phyllite/quartzite unit. Thin bedded greenish gray quartz-mica phyllite and quartzite occupy the stratigraphic interval between the Sangerville and Parkman Hill Formations. This unit was mapped as Waterville Formation by Griffin (1973) and Pankiowskyj and others (1977) and was included in the Sangerville by Ludman (1976).

This unit consists of beds of quartzite from 1 to 8 cm thick alternating with beds of quartz-mica phyllite/schist from 1 to 3 cm thick. The quartzite commonly shows conspicuous crossbeds and small slump structures near the tops of beds. The quartz-mica phyllite/schist contains garnet and staurolite at appropriate metamorphic grades. This unit somewhat resembles the Waterville Formation, but can be distinguished from it by the thicker beds of quartzite and the common internal bedding features.

The unnamed phyllite/quartzite unit has a breadth of outcrop in the vicinity of Skowhegan of approximately 150 m. Because of extensive folding, a thickness of 50 to 70 m may be more realistic.

No fossils have been found in this unit. However, it lies between the Sangerville Formation of Llandoveryan to Wenlockian age and the Parkman Hill Formation of probable Ludlovian age; therefore, its age must be Wenlockian or Ludlovian.

This unit is lithologically and stratigraphically equivalent to the Perry Mountain Formation in western Maine. It has no known correlatives in southern Maine, New Hampshire, Massachusetts or Connecticut.

Parkman Hill Formation. Pankiowskyj and others (1976) described the Parkman Hill Formation as consisting of sulfidic, rusty-weathering metasandstone and metapelite that lies between the Sangerville and Fall Brook Formations. As used in this report it refers to the rusty-weathered metasandstone and metapelite that lies between the unnamed phyllite/quartzite unit described above and the Fall Brook Formation.

As used here the Parkman Hill consists of rusty-weathering, pyritiferous quartz-mica phyllite in beds 0.5 to 10 cm thick. Bedding is difficult to discern on weathered surfaces. Beds of variously rusty, slightly calcareous quartz wacke and graywacke are interbedded with the phyllite, particularly in the western part of the area. The wacke beds have thicknesses up to 15 cm.

Pankiowskyj (1979) estimates a thickness of 200 to 300 m for the Parkman Hill Formation in the Anson quadrangle, but in the vicinity of Canaan (Skowhegan quadrangle) its thickness is only several tens of meters.

Numerous fossil localities have been found in the Parkman Hill Formation (Pankiowskyj and others, 1976). These fossils have been assigned to the range late Llandoveryan to Ludlovian, but the most definitive collections indicate a Ludlovian age.

The Parkman Hill Formation is correlative with the Smalls Falls Formation of western Maine.

Fall Brook Formation. Pankiowskyj and others (1976) used the name Fall Brook Formation for a quartzite sequence in central Maine. Ludman (1977) extended the name to include rocks exposed east of Skowhegan (Figure 1).

The Fall Brook Formation consists dominantly of massively bedded (15 cm to 4 m thick), slightly calcareous quartz wacke. Cross lamination is conspicuous in some beds, and calcareous pods are locally common. Amphibole, diopside, and plagioclase are present at appropriate metamorphic grade. Groups of beds, 1 to several meters thick, of interbedded quartz-mica phyllite/schist and quartzite in beds 1 to 3 cm thick form a subordinate lithology.

No estimates of the thickness of the Fall Brook Formation can be made within the area shown in Figure 1. Pankiowskyj and others (1976) state its thickness to be 1,000 m.

The Fall Brook is unfossiliferous, but because it lies between the rusty unit of Ludlovian age and Early Devonian phyllites, its age is Upper Silurian or Lower Devonian.

The Fall Brook Formation correlates with the Madrid Formation in western Maine. Its correlatives to the south have not been recognized.

Intrusive Rocks. The intrusive rocks are represented by conspicuously foliated "older" granites, dikes of plagioclase granulite and stocks of binary quartz monzonite and biotite granodiorite. The "older" granites are represented by equigranular garnet and biotite bearing granites and pegmatites and by biotite granite containing insets of K feldspar. All varieties are foliated. These "older" granites intrude the Cushing Formation, but do not cut the Silurian turbidite section. These granites have not been dated, but somewhat similar rocks to the northeast have a minimum zircon age of 450 m.y.

Dikes of plagioclase granulite consist of plagioclase, calcite, muscovite, chlorite, and magnetite. These dikes are 60 cm to 3 m thick. They crosscut upright isoclinal folds but are in turn deformed by younger folds. They obviously predate the metamorphism.

Binary quartz monzonite is exposed in four small stocks and lesser bodies (Figure 1). Plagioclase, microcline, quartz, muscovite, biotite, and garnet form an interlocking, inequigranular texture.

Biotite granodiorite is present in large stocks (Figure 1). The biotite granodiorite contains plagioclase, quartz, microcline, biotite, and, locally, garnet or hornblende. It has a perceptible foliation, particularly near the contacts.

The plagioclase dikes, binary quartz monzonite and biotite granodiorite cut Silurian rocks. The plagioclase dikes preceded metamorphism and Ferry (1978) has presented arguments that the quartz monzonite and granodiorite were coeval with regional metamorphism. Several of the stocks have been dated by the Rb/Sr method which indicates ages between 394 and 360 m.y. (Dallmeyer, 1978).

Structural Geology

Structural features include schistosity, cleavage, a variety of lineations, folds, cleavage bands, boudinage, shears, and faults. Fold elements, assigned to three distinct episodes of deformation, consist of late asymmetrical folds, earlier upright folds, and preexisting recumbent folds.

The late asymmetrical folds have axial surfaces that are more or less constant throughout the region (strikes within 10° of north and steep dips). A well-developed cleavage parallels their axial surfaces. These folds deform bedding and schistosity/cleavage of earlier folds; therefore, the plunges and asymmetry of these late folds are variable and depend on the orientation of the surface being folded. No major folds belonging to this

episode have been identified.

These late folds were approximately synchronous with the metamorphism but the deformation continued after the peak of metamorphism. Muscovite and sillimanite have nucleated in the surfaces of cleavage associated with these folds, and porphyroblasts, although they exhibit static growth relative to earlier schistosity, have rotated slightly during the late deformation. Ferry (1978) suggested that the regional metamorphism and emplacement of plutons of granodiorite and quartz monzonite were coeval; this relation would indicate that these folds are Early Devonian in age.

Upright folds are both mesoscopic and major, and these folds control the regional map patterns (Figure 1). In outcrop, these folds have forms that range from relatively open folds to highly flattened isoclinal folds. The openness depends on proximity to competent structural units. Axial surfaces strike northeast and dip steeply; plunges are generally less than 25° , but locally plunges of 45° have been recorded. Regional schistosity and pressure-solution cleavage parallel the axial surfaces of these folds. These upright folds face downward over certain regions and upward over others.

Major upright folds are the obvious structural features as indicated from the distribution of formations (Figure 1). Major folds delineated in Figure 1 are from northwest to southeast: the Athens antiform, the Cornville antiform, the Skowhegan synform, the Hinckley antiform, and the Shawmut antiform. Lesser antiforms and synforms are defined on the basis of the limestone/marble members of the Waterville and Sangerville Formations.

The upright folds predate the metamorphism. The schistosity that is parallel to their axial surfaces is statically enclosed in porphyroblasts, and these folds are cut by dikes (Osberg, 1968) that have been metamorphosed to plagioclase granulites. On the other hand, these folds regionally deform rocks as young as Oriskany and possibly as young as Schoharie. Because the metamorphism has been dated at 400 to 365 million years (Faul and others, 1963; Dallmeyer, 1978), these folds have an Early Devonian age as well.

Evidence of the recumbent folds is mostly indirect: map-pattern, downward-facing upright folds, and stratigraphic relationships. Only at a single outcrop do minor folds display evidence of the earlier recumbent folds. The reconstruction of these folds indicates that they are of Alpine proportions and in south-central Maine face northwest.

The upright folds are interpreted to deform earlier recumbent structures. The Athens, Cornville, Skowhegan, and Hinckley upright folds face upward and are thought to deform the normal limb of a large west-facing recumbent anticline. Near Waterville the Shawmut antiform faces upward and is interpreted to be in the normal limb of a west-facing recumbent syncline. The synform east of the Shawmut anticline and delineated by the limestone member of the Waterville Formation faces downward.

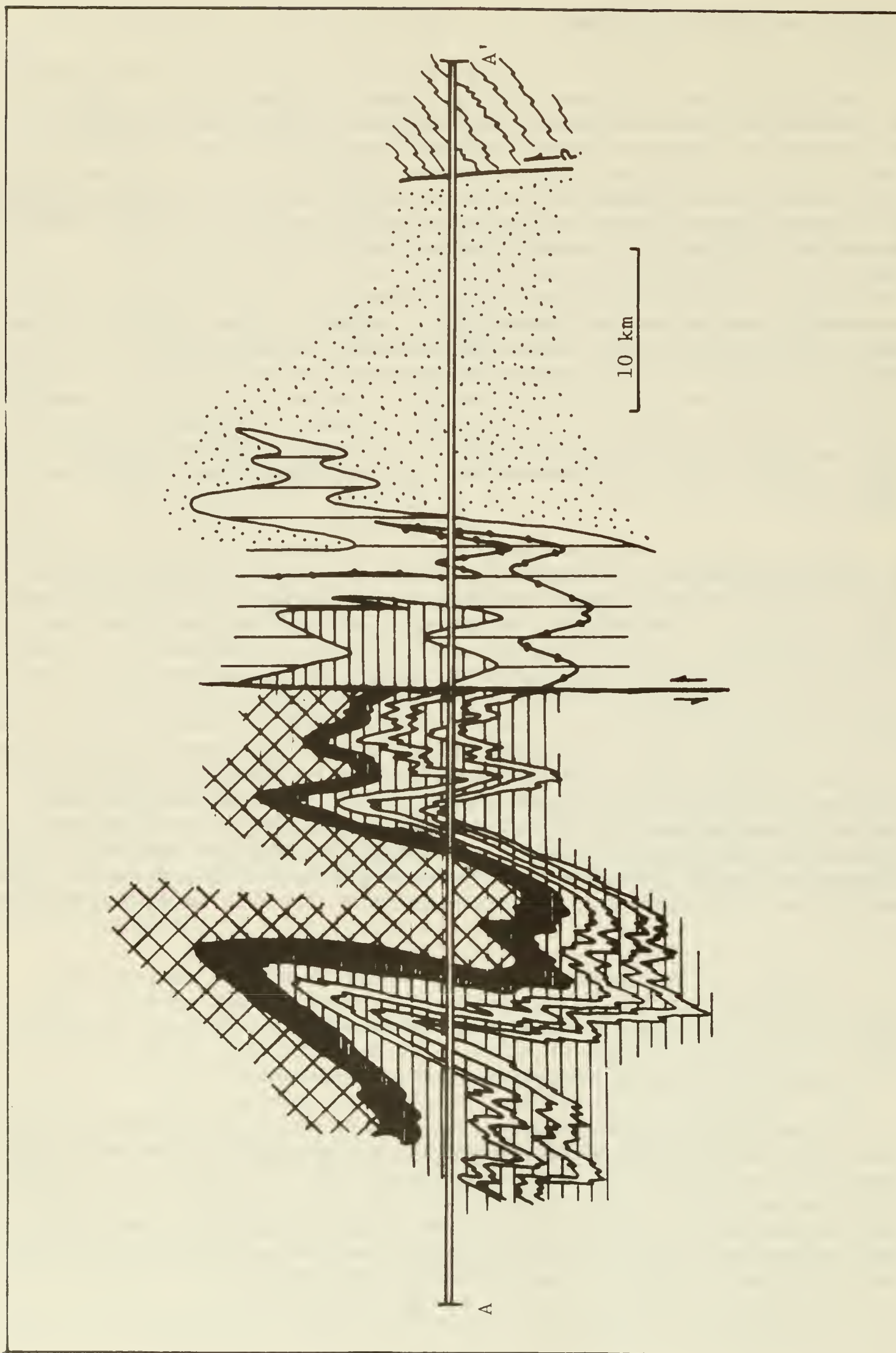


Figure 2. Structure section along A-A'. Patterns same as for Figure 1.

A high-angle fault separates the Waterville Formation from units to the west except west of Waterville where different parts of the Sangerville are juxtaposed (Figure 1). This fault postdates the upright folds and places the Shawmut antiform on the east flank of the Hinckley antiform without an intervening synform. This fault predates the metamorphism.

The contact between the Cushing and Vassalboro Formations is incompletely known (Figure 1). It may be an unconformity or it may be a thrust fault. In either case the contact is folded by the upright folds.

A structure section showing the relationships between the early recumbent folds and the late upright folds is detailed in Figure 2. The line of the structure section is shown in Figure 1.

Metamorphism

Metamorphic rocks of the Silurian turbidite section have been studied by Osberg (1968, 1971, 1974, and unpub. data) and by Ferry (1976a,b). The metamorphism belongs to a Buchan-type facies series. Osberg (1974) mapped isograds in the pelitic rocks on the basis of biotite, garnet, andalusite-cordierite-staurolite, sillimanite, and sillimanite-K feldspar. Ferry (1976a) mapped isograds in the Vassalboro and Waterville Formations on the basis of biotite-chlorite, amphibole-anorthite, zoisite, microcline-amphibole, diopside, and scapolite.

Examination of the chemistry of the associated minerals indicates that the mineral assemblages approached equilibrium near the peak of metamorphism (Osberg, 1971; Ferry, 1978). Estimates of the conditions of metamorphism (Osberg, 1971; Ferry, 1976b) from coexisting minerals indicate that the temperature ranged from $\sim 378^{\circ}\text{C}$ at the biotite isograd to about 550° at the sillimanite isograd. Pressures have been estimated to have been in the range 3,000-3,500 bars. Water pressure was generally less than total pressure.

Dallmeyer (1978) has dated biotite and hornblende from various metamorphic assemblages. Undisturbed $^{40}\text{Ar}/^{39}\text{Ar}$ release spectra and total-gas determinations for biotite produce ages ranging from 335 to 227 m.y.; rocks are younger in the direction of increasing metamorphic grade. Hornblende records older ages than does biotite, and together these ages must be interpreted to reflect diachronous post-metamorphic cooling.

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Itinerary

Mileage

- 0 Assembly point for trip is Stafford's Variety in Athens, Maine (11 miles N of Skowhegan). Starting time is 9:00 A.M.

Stop 1. Outcrop is located on stream 300 m up stream from bridge. Exposures of limestone member of Sangerville Formation.

Gray limestone and sandy limestone in beds 1-13 cm thick. Sandy limestone beds exhibit well preserved cross bedding. Interbeds of siliceous phyllite and biotite quartzite are 1-8 cm thick. Beds strike northeasterly and dip steeply. Well preserved upright, nearly isoclinal folds are preserved in bedding and the cross-bedding indicates that these folds face upward. Isoclinal folds plunge gently south.

Ludman (1977) mapped a fault that was interpreted to isolate this limestone locality. Alternately, the map-pattern of this limestone occurrence may be due to the intersection of the erosion surface with a culmination in a major upright antiform.

Return to cars.

Proceed south on Route 150.

- 1.4 Townline between Athens and Cornville.
- 1.8 Cass Corner. Turn left on West Ridge Road.
- 6.0 Cornville. Continue straight. Hills to east are underlain by biotite granodiorite of Hartland pluton.
- 8.5 Small outcrop of limestone member of Sangerville Formation to right.
- 8.7 Malbons Mills. Turn left.
- 8.8 Stop 2. Park along road. Outcrop is at old mill south of bridge. Exposures of Sangerville Formation.

Under bridge is exposure of quartz-mica phyllite and quartzite in beds 1-3 cm thick. Two thin beds of limestone are intercalated.

At old mill is exposure of massive feldspathic graywacke, polymictic conglomerate, and quartz-mica phyllite. Bedding is 2 cm - 2 m thick. Clasts consist of vein quartz, wacke, feldspar, slate, and volcanic fragments and tend to be flattened in plane of bedding. Maximum size of clasts is approximately 1 cm. Bedding and schistosity are essentially parallel; they strike northeast and dip steeply west.

On east bank of stream exposures are feldspathic graywacke in beds 30-90 cm thick interbedded with 0.5-2.5 cm beds of quartz-mica phyllite. Flame structures at bottom of sandy beds indicate tops east. Late cleavage strikes more northerly than bedding and schistosity and dips steeply west.

Return to cars and reverse direction.

- 8.9 Malbons Mills. Turn left toward Skohegan.

10.7 Stop sign. Junction with Route 2. Turn right onto Route 2.

10.8 Stop 3. Park along Route 2. Exposures of unnamed phyllite/quartzite unit are on promontory west of Great Eddy of Kennebec River. Route 2 is busy highway. Be cautious of traffic!

Quartzite beds 2-7 cm thick show well developed crossbedding. Light gray quartz-mica phyllite, in beds 0.1-3 cm, alternate with beds of quartzite. A few lenses of gray limestone, approximately 30 cm long and 5 cm thick are locally intercalated. Bedding strikes northeast and dips steeply west; crossbeds indicate tops are east. Nearly isoclinal upright folds deform bedding. Crossbeds in the limbs of folds indicate that the folds face upwards. A late cleavage strikes more northerly than bedding and dips steeply northwest.

Return to cars. Reverse direction and proceed east on Route 2.

11.1 Stop 4. Turn right into "picnic area". Exposures of Parkman Hill Formation are located on the bank of the Kennebec River. These outcrops are in west limb of Skowhegan synform.

Rusty-weathering, dark gray pyritiferous quartz-mica phyllite in beds 2-30 cm thick interstratified with beds of rusty-weathered, light purplish gray pyritiferous quartz-feldspar-mica granulite, 8-90 cm thick. Bedding and schistosity strike northeasterly and dip steeply west.

Return to cars and proceed east on Route 2.

12.6 Turn right onto East River Road.

14.5 Stop 5. Park along road. Small exposure of Fall Brook Formation is on north side of road. Good exposures of Fall Brook Formation are not accessible along our route.

Somewhat flinty, purplish gray quartz-plagioclase-biotite-calcite granulite in beds 8-30 cm thick. Medium gray quartz-mica phyllite in layers 1-3 cm thick is interbedded. Quartz veins are numerous. These rocks lie well within the contact aureole of the Hartland pluton. Bedding and schistosity strike northeast and dip steeply west. A small right-hand fold deforms both bedding and a quartz vein; it plunges moderately south. A late cleavage strikes more northerly than bedding and dips steeply west.

Return to cars and proceed east on East River Road.

15.4 Turn left onto Eaton Mountain Road. Outcrops of rusty-weathering, pyritiferous quartz-mica phyllite on left in brook belong to Parkman Hill Formation and are in east limb of Skowhegan synform.

16.7 Eaton Mountain to right is underlain by Parkman Hill Formation and biotite granodiorite of Hartland pluton.

- 18.0 Stop sign. Intersection with Route 2. Turn right and proceed east on Route 2.
- 18.9 Small outcrop of Hartland granodiorite on left.
- 21.5 Canaan. Outcrop of limestone member of Sangerville Formation under bridge.
- 21.8 Route 23 diverges. Continue east on Route 2.
- 22.9 Small exposures of Sangerville Formation on left.
- 23.2 Small outcrops of Sangerville Formation on right.
- 29.4 Junction of Route 152 with Route 2. Turn right on Route 152.
- 29.5 Stop 6. Park on side of road. Exposures of Waterville Formation.

Light purplish and grayish green quartz-mica-chlorite phyllite containing 0.2-8 cm thick beds of quartzite and ankeritic quartzite. Bedding strikes northeasterly and dips steeply east. Upright, isoclinal folds are present at north end of outcrop. These folds plunge gently northeasterly. Foliation in the phyllite wraps around the hinges of these folds.

Return to cars and proceed south on Route 152.

- 32.6 Entering Pittsfield. Stop sign. Turn right to Route 95.
- 33.4 Turn left onto southbound lane of Route 95.
- 35.7 Turn right into rest area. Lunch stop. Continue south on Route 95.
- 44.1 Exposures of Waterville Formation on right.
- 47.5 Bridge over Kennebec River.
- 49.8 Exposures of rusty-weathering, sulfidic quartz-mica phyllite at base of Sangerville Formation.
- 51.0 Bridge over Messalonski Stream.
- 53.1 Turn right off Route 95 using Exit 33.
- 53.4 Stop sign. Turn left onto Oakland Road.
- 54.3 Stop 7. Park at side of highway. Oakland Road is a busy street. Be careful of traffic. Exposure of basal Sangerville Formation is located on north side of highway.

Rusty weathered, dark-gray quartz-mica phyllite interbedded with dark-gray quartzose layers at the east end of the outcrop is the basal unit of the Sangerville Formation. Pyrite is abundant

and crystals have a more-or-less common orientation and do not have associated, quartz-filled pressure shadows. Beds range from 5 mm to 1.5 cm in thickness.

Light-gray, slightly calcareous graywacke and quartz-mica phyllite exposed in the west part of the outcrop are the dominant rocks of the lower unit of the Sangerville Formation. Beds are 15 cm to 50 cm thick and commonly grade from sandy bottoms to phyllitic tops. Sandstone: shale ratios in beds range from 1:1 to 8:1. Small grains of feldspar and shale clasts can be seen at the base of some beds. Good grading can be seen on top of the outcrop and indicates that the cyclically graded unit is younger than the sulfidic phyllite.

Bedding strikes northeasterly and dips steeply. Schistosity cuts bedding at a low angle and the trace of graded beds on schistosity indicate that the section is right-side up. Small shears that offset bedding can be seen on the vertical face, but because the topping direction of bedding does not change across them, the displacement on them is thought to be small. Quartz pods form boudin fillings; other veins are late.

The observed stratigraphic relations indicate that the Sangerville Formation is younger than the Waterville Formation. The outcrop pattern of the Sangerville Formation (Figure 1) in the Shawmut antiform, coupled with the consistent northward plunge of the Shawmut antiform, demands that the Shawmut antiform in part faces downward, i.e., folds the axial surface of an earlier recumbent fold. Because the local section is right-side up, this outcrop must be in the normal limit of the early recumbent syncline (Figure 2).

Return to cars. Continue east on Oakland Road.

- 54.9 Stop light. Continue straight. After crossing bridge, bear right on Grove Street.
- 55.4 Stop sign. Turn left on Water Street.
- 56.2 Stop light. Turn right onto Route 201 and cross bridge over Kennebec River.
- 56.3 Stop light. Turn right remaining on Route 201.
- 56.5 Stop 8. Park in vacant lot to right. Exposure is located over the bank on the Kennebec River. Outcrop is Waterville Formation and its limestone member.

Alternations of light-gray quartz-muscovite-chlorite-phyllite and white to buff quartzite in beds 6 mm to 8 cm thick is a common variant to the Waterville Formation adjacent to the limestone member. Some beds show gradation from quartzite bottoms to pelitic tops. The limestone member consists of gray, slightly micaceous limestone interbedded with rusty, buff-colored quartz-mica.

phyllite. Bedding is 6 mm to 12 cm thick. The contact between the two units is abrupt.

Four areas of the outcrop are particularly worth viewing. In area 1, a northeast-trending isoclinal fold plunges gently to the north. A well-developed cleavage is parallel to its axial surface. Graded bedding near the hinge of this fold has tentatively been interpreted to indicate that these folds face downward. Both limbs of the upright isoclinal fold are cut by cleavage that strikes northerly and dips steeply; right-hand asymmetrical folds that have axial surfaces parallel to the second cleavage are found in both limbs of the upright fold. The concentration of quartz pods marks a shear zone.

At locality 2, an upright, isoclinal synform deforms an earlier isoclinal fold. The upright isoclinal synform has an axial surface that strikes northeasterly and dips steeply; its axis plunges to the northeast. Beds in the hinge of the earlier fold can be traced around the hinge of the later fold, and cleavage that is parallel to the axial surface of the earlier fold can be seen only in its hinge. The plunge of the early fold has not been ascertained. The early fold is thought to relate to the episode of recumbent folding.

At locality 3, beds of the limestone member are folded by isoclinal upright folds. These folds have nearly plane flanks and sharp hinges. Their axial surfaces strike northeasterly and dip steeply, and their axes plunge to the northeast. A second cleavage, best preserved in the beds of phyllite, cuts both limbs of these folds.

At locality 4, a dike of light-gray plagioclase-quartz-muscovite-chlorite-calcite granulite strikes N.33°E. and dips 82°SE. It cuts the bedding at a low angle and has elsewhere been shown to cut the upright isoclinal folds (Osberg, 1968). A faint cleavage oriented parallel to the late, north-trending cleavage cuts the dike. The dike is broken into "rotated" boudinage and the boudin lines plunge steeply. This dike was intruded after the formation of the upright isoclinal folds but before the metamorphism and late deformation.

Return to cars and continue south on Route 201.

- 56.7 Stop light. Continue across bridge on Route 201.
- 56.9 Stop light. Turn left on Route 137.
- 59.4 Stop 9. Turn left onto Pattee Pond Road and park in vacant lot. Exposures of Vassalboro Formation from road cut on Route 137.

Interbedded light-gray quartz-mica phyllite and blue-gray, slightly calcareous quartz wacke are characteristic of the Vassalboro Formation. Possible grading and cross lamination may be seen in the west end of the outcrop.

Open and isoclinal upright folds are identifiable in bedding. Upright folds have axial surfaces that strike northeast and dip steeply. Their axes plunge to the southwest. The "openness" of these folds is controlled by the thick quartzose unit exposed on the south side of the road; away from this unit in less competent lithologies, the folds are more nearly isoclinal. These folds face upward.

Cleavage is parallel to the axial surfaces of the upright folds. In competent beds, it is a pressure-solution cleavage; in the more pelitic beds, it is close-spaced cleavage and micaceous minerals have formed parallel to the cleavage. The cycloidal cross sections of thin quartzite beds interbedded with phyllite may be a pressure-solution effect.

A second set of folds deforms the upright folds and associated cleavage. These folds have axial surfaces that strike northerly and dip steeply. Their axes plunge in various directions depending on the orientations of the surface that they fold. A cleavage parallels the axial surfaces of these folds.

Return to cars. Reverse direction and proceed west toward Winslow on Route 137.

- 61.8 Stop light. Turn left on Route 201 and proceed to Augusta.
- 77.4 Traffic circle east of business district of Augusta. Follow traffic circle 240° and bear right on Routes 9 and 17.
- 77.9 Stop light. Continue straight on Route 9 and proceed to Gardiner.
- 80.6 Outcrop of Vassalboro Formation on left.
- 82.8 Stop light. Turn right onto new bridge over Kennebec River. Town to left is Gardiner.
- 83.0 Stop light. Turn half-left onto Routes 201, 9, and 126.
- 83.3 Stop light. Turn right with Routes 126 and 9.
- 84.2 Bear right with Routes 126 and 9.
- 85.5 Stop 10. Park at side of road. Be careful to traffic. Exposure of Cushing Formation and foliated granite.

Biotite-bearing amphibolite and paragneisses consisting of quartzofeldspathic gneisses and plagioclase gneisses containing garnet and diopside are intruded by "even"-textured, garnet-bearing granite gneiss and granitic gneiss containing insets of K feldspar. Granite and pegmatitic dikes and several orders of veins are present. Biotite reaction zones are present along the margins of several dikes.

The amphibolite, paragneiss, and orthogneiss have been deformed together, and all the rocks have a schistosity that dips moderately toward the northwest. Reclined folds with axial surfaces that dip more gently than the schistosity deform schistosity as well as dikes of orthogneiss, pegmatites, and quartz veins. Boudin lines plunge in approximately the same direction as fold axes. A fold having an axial surface that strikes northerly and dips steeply in the western part of the outcrop is interpreted as a late fold. Enigmatic structures at the middle of the outcrop await an appropriate explanation.

End of trip.

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TRIP COVERAGE

